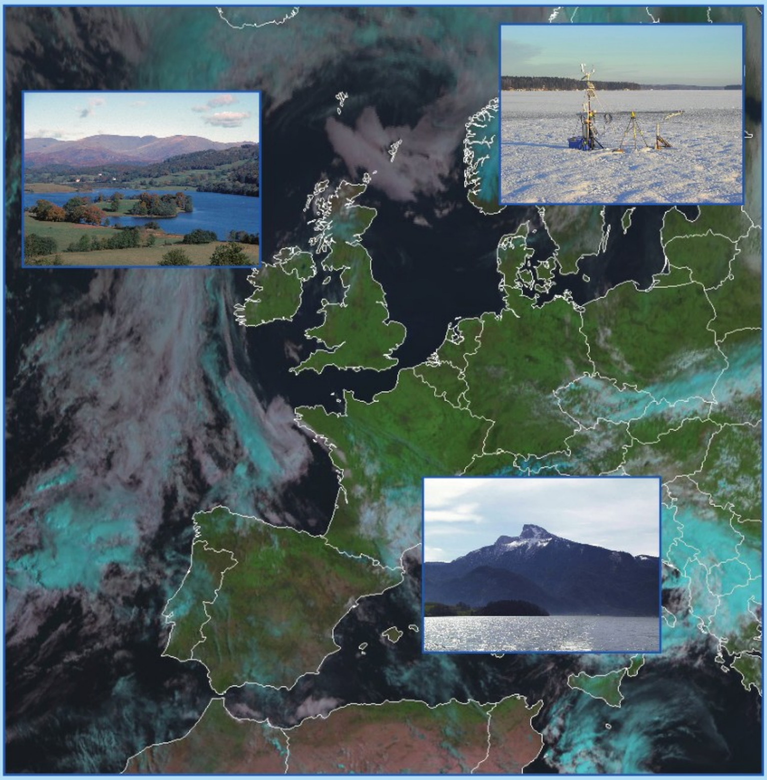
AQUATIC ECOLOGY SERIES

The Impact of Climate Change on European Lakes

Edited by
Glen George



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The Impact of Climate Change on European Lakes

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'Man is born not to solve the problems of the universe, but to find out where the problem applies, and then to restrain himself within the limits of the comprehensible.'

Goethe

I Werin Ewrop
To the Citizens of Europe.

Foreword

Water is a precondition for human, animal and plant life as well as an indispensable resource for the economy. However, the sustainability of this vital resource is currently under threat due to problems such as pollution, the over-exploitation of natural resources, damage to aquatic ecosystems and climate change. Despite the progress made in tackling these problems at a regional and international level, Europe's waters are still in need of increased efforts to get them clean or to keep them clean. Indeed, it is not an exaggeration to say that water protection is one of the major challenges facing the European Union in the new millennium.

The Water Framework Directive (WFD) was adopted precisely for this purpose and establishes a common approach for addressing these problems. To meet the objectives set by the WFD, Member States are required to undertake the measures necessary to ensure that all their surface waters reach the defined 'good ecological and chemical status' by 2015. As this policy has evolved, it has become clear that climate change will have a profound effect on the development of the WFD, an issue covered in the recent European Commission White Paper on 'Adaptation to Climate Change'. For many European regions, water is at the centre of expected impacts of climate change, effects which will be further complicated by changes in water availability and demand.

Research has always played a key role in the implementation of European policy. Water-related research has played and will continue to play an important role in the RTD Framework Programmes of the European Union (EU). Since the late 1980s, water research has been a major component of successive European Commission Framework Programmes (FPs) and has addressed topics that have enhanced our understanding of key processes, their interactions and their role in regulating aquatic systems. These research programmes have also contributed to the development of new technologies and provided tools for the sustainable management of water, as required by the policies implemented by the EU (WFD, Drinking Water, Urban Wastewater, Nitrate and IPPC Directives, etc.). The philosophy of the water-related research funded by the EU Framework Programmes has also evolved over the years to reflect these policy priorities. The earlier Framework Programmes were focused on the extension of basic scientific knowledge. More recent programmes have been designed to support water quality standards and to assist with the development of technologies for 'end-of-pipe' treatments. Over the same period, the scale of the projects supported by the Framework Programmes has also increased. Early programmes were based on a combination of small and medium-size projects that were then clustered to facilitate the integration of the results. These have now been replaced by large-scale Integrated Projects (IP) and Networks of Excellence (NoE) that have the critical mass and the multi-disciplinarity needed to address both the complexity of the problems and the practical needs of stakeholders and end-users. For example, the 5th Framework Programme (1998–2002) included a number of projects that still had a significant emphasis on acquiring new knowledge. In contrast, the 6th Framework Programme (2002–2006) had a strong focus on advancing our understanding of the structure and dynamics of the 'water system' at scales that were relevant to the implementation of policy. These included, enhancing our capability to predict the response of these systems to a combination of pressures and promoting their sustainability through new management concepts, models and guidelines. The new 7th Framework Programme (2007–2013) is even more ambitious in pursuing its role in promoting strong collaborations among research institutes, the academic community and the water industry. A key aim is to develop the most cost-effective, innovative and competitive ways of managing water resources whilst mitigating the effects of climate change on water resources throughout Europe.

The 'Climate and Lake Impacts in Europe' project (CLIME) formed part of the 5th Framework Programme but its multi-disciplinary structure and task integration meant it had much in common with the projects funded by the 6th Framework Programme. In CLIME, climatologists, limnologists and ecosystem modellers from seventeen institutes from nine European countries and the USA combined their resources to investigate the effects of the climate on the dynamics of lakes. The project was funded by the Environment and Sustainable Development research subprogramme of the European Union (EU) and was one of the first EU projects to develop tools and models that could be used to simulate the responses of lakes to both the historical and projected changes in the climate. It is a brilliant example of how the 'European added value', provided by research funded at the EU level, can address highly complex phenomena and analyse patterns of change that can only be understood on a European scale. This book, together with more than a hundred papers already published in specialized journals, demonstrate how these methods and models can now help us to understand and predict the impact of the changing climate on the dynamics of lakes.

The catchment based approach adopted by CLIME very much reflects the operational requirements of the WFD. The results summarized in this book will consequently be of interest to those charged with the regulation and the management of water resources in Europe. The model simulations, in particular, provide the sector with the information they need for policy making and implementation. The models have been tested at sites exposed to the very different weather patterns experienced in Northern, Western and Central Europe. The breadth of environmental conditions covered by CLIME, coupled with the probabilistic approach used in the Decision Support System (CLIME-DSS), mean that the projections are robust enough to be utilized by water regulators and managers in parts of Europe not covered by the project.

A key element in all projects funded by the EU is the method used to ensure the effective dissemination and exploitation of the results. These methods must demonstrate the benefits gained by European society in ways that are readily apparent to the tax-payers, the ultimate sponsors of the work. In this respect, I would like to thank and congratulate all the CLIME partners for the exemplary way in which they have disseminated the results of their researches through the publication of this book. I very much hope that it will get the reception it deserves from water companies, policy makers, regulatory authorities and the scientific community. The former will, hopefully, use the CLIME results to help them assess the potential impact of climate change on the strategies employed to safeguard our water resources. It is the feedback from the application of the results presented here that will be most welcomed by the CLIME consortium. It is this feedback, above all else, that will have made the effort involved in producing this book a most rewarding and worthwhile exercise.

Christos Fragakis
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Chapter 1
The Impact of Climate Change on European Lakes

Glen George

1.1 Introduction

Llym awel, llwm bryn	*Keen wind on the bare hill*
Anodd caffael clyd	*It is difficult to find shelter*
Llygra rhyd, rhewid llyn	*The ford is polluted and the lake frozen*

The above quotation, taken from a thirteenth century Welsh manuscript (Jones and Jarman, 1982), elegantly expresses the essence of this book. In a few words, the poet encapsulates the practical consequences of an extreme climatic event and describes its impact on the quality of water in a river and the physical characteristics of a lake. This poem was probably composed at the beginning of the thirteenth century when Britain experienced some very severe winters (Ogilvie and Farmer, 1997). Today, severe winters are rare but we are becoming increasingly concerned about the impact of mild winters and warm summers on the quality of the water in our lakes and rivers. The ecological status of many lakes in Europe has changed dramatically over the last 20 years. Many of these changes are the result of anthropogenic influences in the catchment but some are also driven by changes in the regional climate (George, 2002; Blenckner and Chen, 2003; Straile et al., 2003). In Northern European lakes, the most important climatic effects are those associated with the extension in the duration of the ice-free period (Weyhenmeyer et al., 1999; Weyhenmeyer et al., 2005). In Western European lakes, the most important impacts are those connected with the increased winter rainfall (George et al., 2004) and the changing frequency of calm summer days (George et al., 2007).

The 'Climate and Lake Impacts in Europe' project (CLIME) was co-funded by the European Commission's Directorate General Research and formed part of the 'Catchmod' cluster of projects on the integrated management of water basins (Blind et al., 2005). The primary objective of CLIME was to develop models that could be used to simulate the responses of lakes to future, as well as past, changes in the

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climate. These models were validated using long-term data acquired from a number of sites and perturbed with the outputs from the latest generation of Regional Climate Models (Räisänen et al., 2004). The secondary objective of CLIME was to analyze the inter-annual variations observed in a range of different lakes and relate these variations to local changes in the catchment and regional changes in the weather. CLIME was a 3-year project that ended in December 2006. Since then, a number of presentations have been delivered at international conferences and more than a hundred papers published in specialized journals. In this volume, we present an overview of these results and explain how analyses of this kind can be used to support climate impact studies on other aquatic systems.

1.2 Climate Change and the Water Framework Directive

The CLIME project was designed to provide strategic support for the Water Framework Directive (WFD). The WFD represents a fundamental change in the way water resources are managed in Europe. It establishes a single system of assessment that deals with all aquatic systems in a holistic way. The guiding principle of the directive is the achievement of 'good ecological status' for all ground and surface waters by 2015. Surface waters are defined as having a good ecological status if their physical, chemical and biological characteristics match those of similar systems exposed to minimal anthropogenic influences. When this scheme was developed, climate change was not the pressing issue it is today. In future, the procedures used to support the directive will have to be modified to accommodate both the direct and indirect effects of the changing climate. In the current scheme, catchment managers are required to define the 'natural variability' associated with lakes located in different parts of Europe. This variability is projected to increase as the world becomes warmer (Temnerud and Weyhenmeyer, 2008) so the criteria used to define their ecological status will have to be changed to accommodate this uncertainty. A number of reports on the potential impact of climate change on the WFD have already been produced (Arnell, 2001; Wilby et al., 2006; Eisenreich et al., 2005). CLIME was one of the first projects to explore the effects of the projected changes in climate on water quality as well as quantity. In 2005, it was succeeded by another EU project (Eurolimpacs) that was designed to evaluate the effects of global warming on rivers and wetlands as well as lakes (www.eurolimpacs.ucl.ac.uk).

For the regulatory authorities, the key issue is to distinguish those lakes where the deterioration in water quality can be related to changes in the management of the catchment from those where the main driver is the regional change in the climate. In practical terms, this means identifying those sites where the remedial measures would be most effective before setting new targets for the less resilient sites. The schematic in Fig. 1.1 shows how the water quality targets set for a particular lake might have to change to accommodate the new boundary conditions imposed by the changing climate. When these boundary conditions are fixed (Fig. 1.1a), the measures adopted to achieve 'good ecological status' may remain effective for

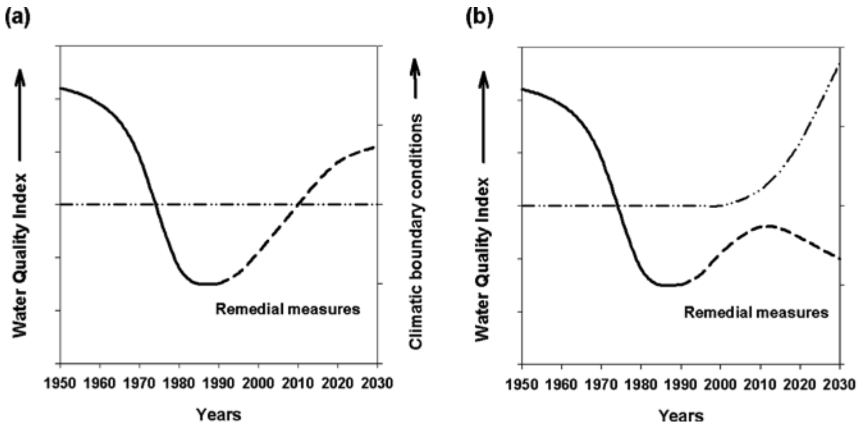

Fig. 1.1 Schematic showing the effects of introducing remedial measures to improve water quality: (**a**) Where there has been no systematic change in the local climate. (**b**) Where there has been a progressive change in the local climate

some considerable time. When the boundary conditions change (Fig. 1.1b), the remedial measures are 'chasing a moving target' and the catchment managers may have to revise the objectives set for the lake.

A good example of such a 'climate-driven' response is that provided by Lake Erken in Sweden (59° 25' N; 18° 15' E). Lake Erken is a mesotrophic lake that used to be covered with ice for at least four months in the year. Water samples for nutrient analysis have been collected from this lake at regular intervals since the 1960s and show no evidence of a significant increase in the nutrient load (Chapter 8, this volume). The recent extension in the ice-free period has, however, intensified the recycling of phosphorus (Chapter 15, this volume) and led to a sustained increase in the summer biomass of phytoplankton. Figure 1.2 shows the inter-annual variation in the chlorophyll concentrations recorded in the lake for two different periods. In the late 1970s (Fig. 1.2a), when the lake was free of ice for about 230 days each year, the average summer concentration of chlorophyll was 2.6 μg L^{-1}. In the late 1990s (Fig. 1.2b), when the ice-free period had increased to around 260 days, the average summer concentration of chlorophyll was 4.4 μg L^{-1}. The factors responsible for this increase were the extension of the growing season, the increased consumption of oxygen in deep water and the enhanced recycling of phosphorus from the sediment. These changes also had a major effect on the qualitative composition and seasonal dynamics of the phytoplankton. In the 1980s, the spring diatom bloom appeared earlier in the year than in the 1970s and there was a five fold increase in the summer abundance of the blue-green alga *Gloeotrichia* (Weyhenmeyer et al., 1999). Results of this kind demonstrate that small changes in the physical characteristics of a lake can have a disproportionate effect on its chemistry and biology. Here, an increase of a few weeks in the duration of the ice-free period resulted in changes that mimicked

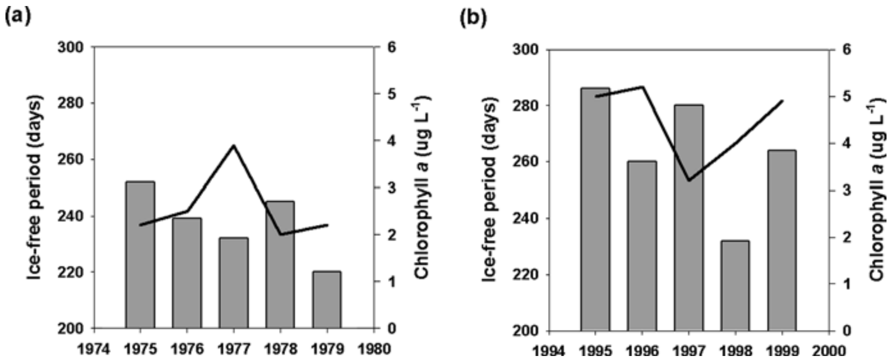

Fig. 1.2 The effect of an extension in the ice-free period on the summer biomass of phytoplankton in Lake Erken: (**a**) The condition of the lake in the 1970s. (**b**) The condition of the lake in the 1990s. The *bars* show the ice-free period and the *lines* the biomass of phytoplankton (modified from George, 2002)

those often associated with cultural eutrophication. Freshwater ecologists frequently underestimate the importance of time as a controlling factor. Most bloom-forming species of algae grow very slowly, so an extra cell division in early summer can lead to a substantial increase in the biomass present much later in the year.

1.3 The Impact of Regional Variations in the Weather on Water Quality

Some water quality problems can be directly related to systematic variations in the local weather. A good example is the enhanced growth of cyanobacteria in warm, calm summers. Such ‘blooms’ were once considered to be an inevitable consequence of eutrophication but changes in the weather also play a major part in their seasonal development (Paerl and Huisman, 2008). Cyanobacteria grow slowly so factors such as a sustained reduction in the flushing rate and changes in the intensity of thermal stratification can have a profound effect on their relative abundance. Some species also produce gas vacuoles that allow the cells to float to the surface and accumulate downwind when wind speeds are low and there is little turbulent mixing (George, 1992).

Figure 1.3 shows the extent to which the reported frequency of algal blooms in England and Wales in 1989 matched the spatial variation in the average summer wind speed. The map in Fig. 1.3a is based on the results of a questionnaire circulated to water managers by the National Rivers Authority (NRA, 1990). The wind speed map in Fig. 1.3b is a simplified version of that given in Hulme and Barrow (1997). In 1989, a very mild winter was followed by an unusually warm, calm summer. Under these conditions, blooms of cyanobacteria were recorded in a number of lakes and reservoirs but the highest frequencies were noted in inland areas where the wind-speeds were lower. In a warmer world, the incidence of such blooms is

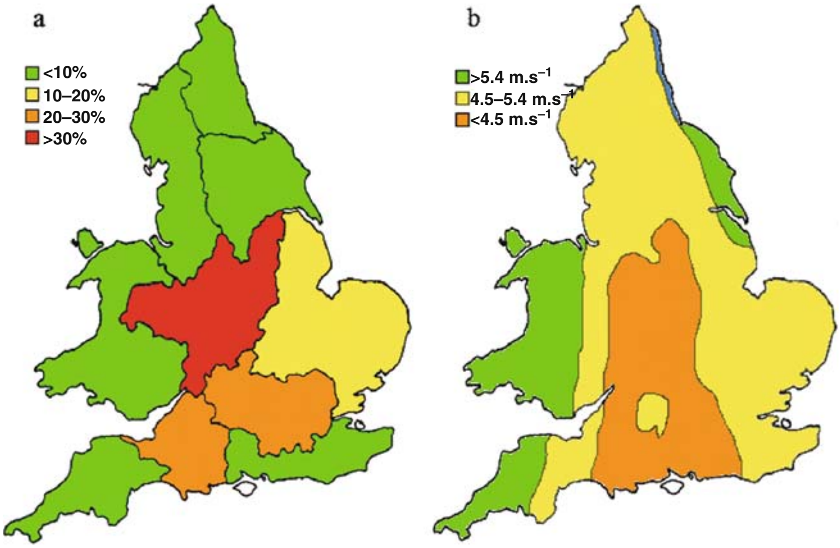

Figure 1.3 (**a**) The incidence of cyanobacteria blooms in England and Wales in 1989. (**b**) The spatial variation in the average summer wind speed recorded in England and Wales between 1961 and 1990 (modified from Hulme and Barrow, 1997)

likely to increase since the cyanobacteria will appear earlier in the year and their growth will be enhanced by more prolonged periods of stable stratification. Serious cyanobacterial blooms were reported in a number of European lakes during the warm summer of 2003 and there is evidence to suggest that some species have already extended their geographical range (Wiedner et al., 2007).

1.4 The Organization of the Project

In CLIME, seventeen partners from ten countries combined their resources to explore the direct and indirect effects of the projected changes in the climate on the dynamics of lakes. The map in Fig. 1.4 shows the geographic distribution of the project partners. Fifteen partners were based in countries that are members of the European Union; one partner was from Switzerland and one from the USA.

The organization of the project was designed to break down the barriers that often exist between climatologists, limnologists and ecosystem modellers. The schematic in Fig. 1.5 shows the key components of the project: ‘Projection’, ‘Attribution’, ‘Simulation’, ‘Prediction’ and ‘Implication’.

In the ‘Projection’ component, climatologists from Sweden and the UK used the results from two Regional Climate Models and two greenhouse gas emission scenarios to produce the projections required by the analysts and modellers. In the ‘Attribution’ component, the data acquired from the selected lakes was processed and the observed inter-annual variations related to local changes in the catchment

Fig. 1.4 The location of the seventeen partners in the CLIME project. *Austria*: OEAW – Austrian Academy of Sciences, Institute for Limnology; UIBK – University of Innsbruck, Institute of Zoology and Limnology. *Estonia*: EAU – Estonian Agricultural University, Institute of Zoology and Botany. *Finland*: UH – University of Helsinki; TKK – Helsinki University of Technology, Water Resources Laboratory. *Germany*: FBV – Institute of Freshwater Ecology and Inland Fisheries, Berlin; UKON – University of Konstanz, Limnological Institute. *Hungary*: TUBID – Budapest University of Technology and Economics; UVES – University of Veszprém. *Ireland*: MI – Marine Institute, Galway; TCD – Trinity College Dublin. *Sweden*: SMHI – Swedish Meteorological and Hydrological Institute; UU – University of Uppsala, Erken Laboratory. *Switzerland*: EAWAG – Swiss Federal Institute of Environmental Science and Technology. *UK*: CEH – Natural Environment Research Council, Centre of Ecology and Hydrology; UEA – University of East Anglia. *USA*: New York City Department of Environmental Protection, Bureau of Water Supply

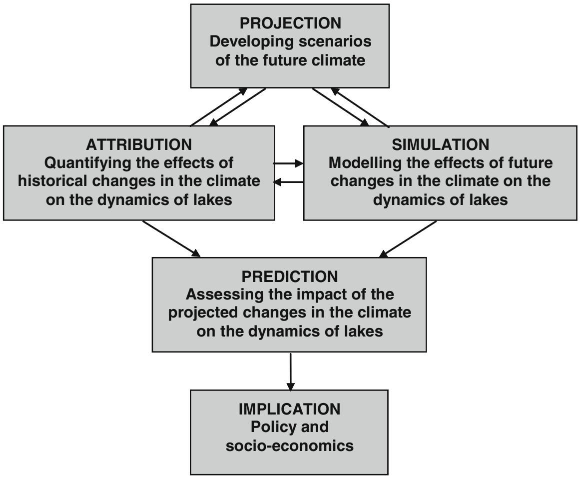

Fig. 1.5 The five components of the CLIME project. Each component included a number of Workpackages with defined objectives and a range of 'deliverables'

and regional changes in the weather. In the 'Simulation' component the modellers used a coupled series of climate, catchment and lake models to explore the potential effects of changes in the climate on the dynamics of lakes. The same models were used to quantify the changes expected in the three European regions and a stochastic approach used to establish the uncertainties associated with the individual water quality simulations. The 'Prediction' component used a Decision Support System (CLIME-DSS) to assimilate the data acquired by the analysts and modellers and to display the results in a form that was accessible to a non-expert end-user. The CLIME-DSS and its source code are now in the public domain and can be downloaded from: http://geoinformatics.tkk.fi/twiki/bin/view/Main/CLIMEDSS. The final 'Implication' component addressed some of the practical consequences of the projected changes in the climate. These included: a socio-economic analysis of the perceived risks, a strategic analysis of the consequences for the water industry and an assessment of the impact of climate change on the implementation of the Water Framework Directive.

1.5 The Scope of the Book

The scope of the book reflects the objectives outlined in a 'Description of Work' submitted to the European Commission in 2002. Most of the results are based on work completed between 2003 and 2006 but a few examples have been included from an earlier EU funded project (REFLECT). The lakes and catchments included in CLIME represent some of the most intensively studied sites in Europe. All have been monitored at weekly or fortnightly intervals for at least 20 years and some have records that extend into the 1930s and 1940s. Table 1.1 lists the main CLIME sites arranged in descending order of latitude and notes some key characteristics.

Table 1.1 The characteristics of the lakes included in CLIME. 'P' denotes a Primary Site that was subject to particularly intensive study and 'S' a supporting Secondary Site

Name	Latitude	Location	Status	Surface area (km^2)	Max. depth (m)	Trophic status
Valkea-Kotinen	61.14°N	Finland	S	0.04	6.5	Dystrophic
Pääjärvi	61.04°N	Finland	P	14	85	Mesotrophic
Erken	59.84°N	Sweden	P	28	21	Mesotrophic
Mälaren (Galten)	59.45°N	Sweden	S	64	19	Eutrophic
Peipsi	58.36°N	Estonia	S	3,555	15.3	Eutrophic
Võrtsjörv	58.08°N	Estonia	P	270	6	Eutrophic
Müggelsee	52.26°N	Germany	S	7	8	Hypereutrophic
Windermere (N)	54.24°N	UK	S	8	60	Mesotrophic
Esthwaite Water	54.21°N	UK	P	1	15	Eutrophic
Lough Feeagh	53.50°N	Ireland	P	4	45	Oligotrophic
Lough Leane	52.00°N	Ireland	S	20	60	Mesotrophic
Mondsee	47.83°N	Austria	P	14	68	Mesotrophic
Lake Constance	47.67°N	Germany	P	94	254	Mesotrophic
Greifensee	47.35°N	Switzerland	P	9	32	Eutrophic
Lake Zurich	47.27°N	Switzerland	S	67	136	Mesotrophic
Piburgersee	47.17°N	Austria	S	13	25	Oligotrophic
Lake Balaton	46.92°N	Hungary	P	593	12	Eutrophic

The period used for our historical analyses was the 1961–1990 reference period recommended by the Intergovernmental Panel on Climate Change (IPCC, 2001). Particular attention was paid to the climatic sensitivity of different types of lakes and to the atmospheric features that influenced their seasonal dynamics.

The most northerly lake was Valkea-Kotinen, a dystrophic lake situated in a natural forest in southern Finland. The most southerly lake was Balaton, a large eutrophic lake on the Hungarian plain. The sites designated as Primary Sites (P) were used to calibrate the models and analyze long-term trends. Those designated as Secondary Sites (S) were used for the regional assessments and the supra-regional analyses of the factors influencing long-term change. Most of the lakes were deep and remained thermally stratified throughout the summer but Peipsi, Võrtsjärv and Balaton were shallow and relatively well mixed. All the lakes were subject to a variety of anthropogenic influences ranging from upland grazing (Lough Feeagh) to intensive agriculture and tourism (Lake Balaton). In the long-term analyses, we used simple statistical techniques to separate the effects associated with local changes in the catchment from those driven by the regional variations in the climate.

Water quality problems are usually addressed on a site-by-site, issue-by-issue basis. A central aim of CLIME was to move away from the 'my lake', 'our catchment' approach to compare the patterns and processes observed on a pan-European scale. An important element in this strategy was the time devoted to analyzing the long-term records. These records provide an effective means of assessing the climatic sensitivity of the different lakes and quantifying the impact of the more extreme variations in the weather. Many of these long-term monitoring programmes are currently struggling to survive. Several freshwater research groups in Europe have recently been downsized and resources allocated to more ‘fashionable’ topics. If this trend continues, we will be left with records that are too incomplete for systematic analysis, acquired at frequencies that do not match the accelerating pace of change. In CLIME, we also established a network of automatic monitoring stations to record the responses of the lakes to very short-term changes in the weather (Rouen et al., 2005). Some of these stations have now been included in the GLEON network (www.GLEON.org) but securing long-term funding for such installations is very difficult. The other important element in our research strategy was the integration of the individual modelling activities. In most large projects, modelling studies of this kind are conducted by specialists who then circulate the results to other members of the consortium. In CLIME, we adopted a different approach, where representatives from each site were trained to run their own simulations. In this way, many functional aspects of the models were improved and modelling skills acquired by new groups within the European research community.

1.6 The Organization of the Book

The chapters in the book are arranged in a way that emphasizes the link between the description and the simulation of environmental change. Chapter 1 explains the aims of CLIME and describes its organization and management. Chapters 2 and 3 introduce the models that formed the core of the CLIME project. In Chapter 2, Samuelsson describes the Regional Climate Models used and explains the methods adopted to extrapolate these results to a catchment scale. In Chapter 3, Schneiderman et al. explain the rationale for using the Generalised Watershed Loading Function (GWLF) model to quantify the effect of changes in the climate on the flux of nutrients. Chapters 4 and 5 consider the effects of the changing climate on the phenology and dynamics of lake ice. The historical analyses presented by Livingstone et al. in Chapter 4 use data collated from a range of lakes to summarise the phenology of ice in different parts of Europe. In Chapter 5, Leppäranta uses a numerical model to simulate the growth and decay of ice on a lake in southern Finland. This model is new and includes procedures that simulate the development of the different layers within the ice sheet. Chapters 6 and 7 consider the effects of the changing climate on the temperature and mixing characteristics of lakes. Chapter 6, by Arvola et al., documents the long-term changes in temperature observed in more than twenty European lakes. Chapter 7, by Jones et al., uses a 1-D model to compare the thermal responses of lakes situated in the UK, Sweden and Austria. Chapters 8 and 9 investigate the effects of the changing climate on the supply and recycling of phosphate. In Chapter 8, Pettersson et al. use a case study approach to describe the historical variations in the phosphorus content of lakes situated in Sweden, Estonia and the UK. In Chapter 9, Pearson et al. explain how the GWLF model was used to quantify the impact of the projected changes in the climate on flux of phosphorus in lakes located in Finland, Sweden, Estonia and the UK. Chapters 10 and 11 consider the effects of the changing climate on the seasonal variations in the concentration of nitrate. In Chapter 10, George et al. use a statistical approach to identify the climatic variables regulating the flux of nitrate in eight European lakes. The modelling studies described by Moore et al. (Chapter 11) then use the GWLF model to quantify the effects of changes in the climate on the export of dissolved inorganic nitrogen from catchments in Finland, Sweden, Estonia, the UK and Ireland. The observational and modelling studies described in Chapters 12 and 13 examine the factors influencing the supply of dissolved organic carbon (DOC) to lakes in Finland, Sweden, Estonia, the UK and Ireland. In Chapter 12, Jennings et al. describe some of the factors that have contributed to the recent increase in the flux of DOC. The effects of future changes in the climate on the export of DOC are then explored by the modelling studies described by Naden et al. in Chapter 13. In Chapter 14, Nõges et al., use long-term records to compare the climatic responses of phytoplankton in lakes distributed throughout the three regions. In Chapter 15, Blenckner et al. show how the results from different simulation models can be combined to study the effects of the projected changes in the climate on the flux of nutrients and the growth of phytoplankton. Chapters 16 and 17 describe the ways in which changes in the circulation of the atmosphere influence the inter-annual variations observed in a number of CLIME lakes. In Chapter 16, George et al. use the 'weather typing' approach to highlight the impact of different pressure patterns on the surface temperature of lakes. The statistical analyses presented by Livingstone et al. (Chapter 17) show that limnological time-series collated from a range of different lakes frequently display a high degree of spatial coherence. Chapters 18, 19 and 20 review some of the more important climate-related changes observed in the CLIME lakes over the past forty years. In Northern Europe, Blenckner et al. (Chapter 18) report that the most important effects were those associated with the extension of the ice-free season. In Western Europe, George et al. (Chapter 19) explain that the most important effects were connected with the flushing action of the rain and the mixing effect of the wind. The effects of the changing climate on the lakes of Central Europe are reviewed by Dokulil et al. in Chapter 20. They use data collated from lakes located in seven countries to describe the observed variations and relate these to global-scale changes in the climate. The most innovative development in CLIME was the Decision Support System described by Jolma et al. in Chapter 21. The CLIME-DSS is a web-based application that allows the user to visualize the projected changes in the European climate and explore their effect on the dynamics of lakes. The Bayesian Networks (BNs) used to drive the CLIME-DSS were based on the analysis of long-term data from a number of different lakes and the results from a large number of model simulations. In Chapter 22, Bateman and Georgiou summarize the results of the socio-economic studies commissioned by CLIME. In these studies, they assessed the public's awareness of climate-related problems in the water industry and quantified their willingness to pay for remedial measures. Chapter 23, by Frisk and George, discusses some of the ways in which the approaches developed in CLIME can be used to support the future development of the Water Framework Directive. In

Chapter 24, Janus uses a case-study approach to demonstrate some of the potential effects of climate change on the supply of potable water. The examples are taken from a network of reservoirs maintained by New York City where climate, lake and catchment models are being used to assess the risks associated with global warming.

1.7 Concluding Remarks

CLIME was conceived as an integrated project where specialists from seventeen partners complete a sequence of inter-related tasks. Administrative tasks were kept to a minimum and several weeks in each year devoted to ‘hands-on’ workshops on model building and data processing. From the outset, we were fortunate in being able to access the results from the latest generation of Regional Climate Models and downscale these to catchment scales. This would not have been possible without the active participation of climatologists from the Swedish Meteorological and Hydrological Institute and the Climate Research Unit in the UK. In CLIME, as in many other investigations, the catchment approach advocated by Likens (2001, 2004) provided the framework needed to integrate the results. By the end of the project, we had analysed long-term records from more than thirty lakes and performed more than 15,000 individual model simulations. The decision support system (DSS) developed by our colleagues from Finland, provided the key to assimilating all this information and displaying the results in a form that was readily accessible to end-users in the water industry.

Managing large European projects, like CLIME, is a challenging task. The key to success is the personal chemistry that develops within the group as the project evolves. In an article on the limitations to intellectual progress in ecosystem science, Likens (1998) described the personal qualities that characterize an effective team. They include: a readiness to share both ideas and data and a commitment to work in an open and flexible way. These qualities are difficult to sustain in a competitive world, but the projects funded by the Commission help to sustain this ethos within the European research community.

Since CLIME had to be completed in three years, particular attention had to be paid to inter-dependencies in the project. Much of the detailed planning was devolved to Special Interest Groups that met twice a year to review progress and plan future work. The other group that played a key role in managing the project was the Scientific Advisory Board. The ‘core’ members of the Board were Karin Pachel (Estonian Ministry of the Environment), Per Eriksssson (Norrvatn – Northern Water Board, Sweden), Alastair Ferguson (Environment Agency, England and Wales), Alfred Jagsch (Institute for Water Ecology, Fishery Biology and Lake Research, Austria), Gabor Molnar (Lake Balaton Development Co-ordination Agency, Hungary), Pius Nierderhauser (Department of Water Protection, Zurich Canton Bureau of Waste, Water, Energy and Air, Zurich, Switzerland) and Lorraine Janus (New York City Department of Environmental Protection). These meetings were ably chaired by Tom Frisk (Pirkammaa Regional Environment Centre, Finland) who also attended a number of *ad hoc* meetings organized to evaluate the performance of the Decision Support System.

This synthesis volume was planned in the opening stages of CLIME but all the chapters were written after the project had officially ended. I extend my thanks to all the contributors whilst acknowledging the pivotal role played by the first authors of the individual chapters. The quality of the final drafts owes much to the constructive comments of forty six international reviewers. Finding independent reviewers for these large European projects can be difficult, so I am very grateful to all those who responded so positively to my pleas for help.

I would like to thank my colleagues from the Freshwater Biological Association (Windermere) and the Centre for Ecology and Hydrology (Lancaster) for their support over the years. The long-term data for the English lakes are held jointly by the Freshwater Biological Association and the Centre for Ecology and Hydrology (CEH). I am also grateful to Diane Hewitt and Fiona Carse for help with data processing and Margaret Hurley for statistical advice. The automatic monitoring stations used in CLIME were designed by Martin Rouen from CEH (now at Lakeland Instrumentation) with support from Jack Kelly, Paul Jones, Paul Hodgson and David Benham from CEH. Special thanks are due to my colleague Stephen Maberly for his support at a difficult time for the group at Lancaster. A number of administrative staff from CEH Windermere and CEH Lancaster also helped with the management of the project. These included Undine Day, Yvonne Dickens, Helen Grosse, Tracy Pow, Katherine Fawcett and Gaynor Greenwood. The cost of producing the index and the colour illustrations in Chapters 1 and 19 was covered by a special grant from CEH.

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Chapter 2
Using Regional Climate Models to Quantify the Impact of Climate Change on Lakes

Patrick Samuelsson

2.1 Introduction

The world's climate is changing at an unprecedented rate. Globally–averaged temperatures reveal that thirteen of the last 14 years (1995–2008) were amongst the warmest on record. The rate of warming over the last 50 years is nearly twice that recorded for the last hundred years (MICE, 2005). In Europe, significant increases in precipitation have been recorded in most northern areas and there has been a corresponding decrease in the area of seasonally frozen ground. The most recent report produced by the Intergovernmental Panel on Climate Change (IPCC, 2007) confirms that mankind is responsible for much of this change and notes that the evidence for this anthropogenic effect is now overwhelming. Stern (2006) published a report on the economic consequences of climate change which concluded that 'the costs of stabilising the climate are significant but manageable; delay would be dangerous and much more costly'.

The climatic changes observed in recent decades have already had a major effect on rivers and lakes throughout Europe. Chapters 18, 19 and 20 in this volume summarise some important impacts on the dynamics of lakes whilst Chapters 23 discusses the effect of these changes on the implementation of the Water Framework Directive. This chapter starts with a discussion of the usefulness and limitations of climate modelling. It then discusses the strengths and weaknesses of dynamical and statistical downscaling and explains how these procedures were used to support the different modelling tasks in CLIME. Outputs from the selected Regional Climate Models (RCMs) was circulated to all partners in the first year of the project and then used to inform the historical analyses and drive the model simulations. In the final year of CLIME, RCM outputs for the European area were used to provide the climatic inputs required by the Decision Support System (CLIME-DSS) described in Chapter 21. The CLIME-DSS is a web-based application that allows the user to

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visualize the climatic changes projected for the European area and explore some possible effects on the dynamics of lakes. The DSS is already in the public domain (http://geoinformatics.tkk.fi/bin/view/Main/CLIMEDSS).

The chapter concludes by summarizing the climate projections produced for the three European regions and describes some of the new RCM products that are now becoming available.

2.2 The Role of Climate Modelling in Impact Assessments

Decision makers in the European Union urgently need detailed projections of the future climate and information on the uncertainties associated with these projections. Climate models represent an important tool in the understanding of the earth's climate system. For the current and historical periods, they complement the observational methods used to analyse the climate. For future periods, they are the only tools available to assess the rates of change that can be expected under different emission scenarios. Before they can be used to assess the possible future status of the climate, the climate models must be able to reproduce the essential features of the historical climate. However, good agreement with the observed climate does not necessarily mean that the models will perform well outside the control period. This reflects the uncertainties that still exist regarding key physical processes and the way in which these are parameterized in the models. Our knowledge on how the ocean-atmosphere system works continues to improve but many uncertainties still remain and have a major effect on the results produced by the different models.

One question frequently posed by potential end-users is: ‘Are climate models reliable?’ The usual response is to say ‘yes’, or at least reliable enough to be used as tools to test the climatic sensitivity of the earth and its ecosystems. A good illustration of such a sensitivity analysis is that shown in Fig. 2.1. These simulations were first presented by Stott et al. (2000) and are now widely used to demonstrate the impact of anthropogenic factors on the global climate (Hegerl et al., 2007). Figure 2.1a shows the results of some simulations where an ensemble of General Circulation Models (GCMs) were run for a historical period using only natural forcing. The simulated temperature did not show any increasing in the second half of the 20th century, a situation that is in stark contrast to the historical observations. Figure 2.1b shows the result from another ensemble where the natural forcing was held constant but the forcing due to the increased concentration of greenhouse gases was adjusted according to the most recent estimates. These simulations showed a marked increase in the temperatures estimated for the end of the 20th century, but the warming observed in the early part of the century was not captured. Figure 2.1c shows the result when both sets of factors – natural and human – were included in the ensemble simulations. When the two factors are combined there is a good match between the projections and the observations which gives us some confidence in the performance of the model and our understanding of the key drivers.

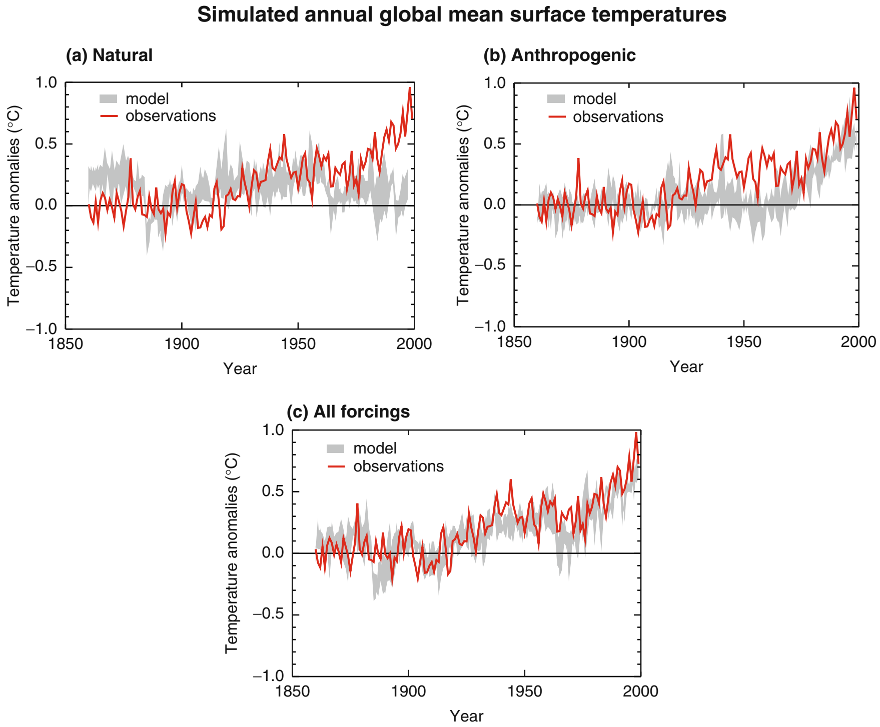

Fig. 2.1 The observed changes in globally-averaged temperatures (1850–2000) compared to those simulated by an ensemble of simulations from HadCM3 perturbed by: (**a**) Natural forcing. (**b**) Anthropogenic forcing. (**c**) Both natural and anthropogenic forcing. Reprinted from IPCC (2001)

2.3 Dynamical and Statistical Downscaling

General Circulation Models (GCMs) operate on quite a coarse spatial scale when compared to the scale of most environmental impact assessments. This limitation is particularly important where the area of interest is topographically complex or where the parameters of interest have steep horizontal gradients. For example, the horizontal distribution of precipitation is very sensitive to small scaled inhomogeneities in the landscape. Also, the larger the grid squares used in the simulation the stronger the tendency to distribute precipitation over a large number of low-intensity events. Thus, if precipitation is an important element in the assessment it is almost always necessary to downscale the output generated by a GCM.

There are basically two methods available for downscaling climate projections; the dynamic method using regional climate models (RCMs) and the statistical method that relies on a more empirical approach. The main strengths of the RCMs are that they respond in a physically consistent way to different forcings and include most of the feedbacks that exist in the climate system. They are, however, very dependent on the boundary conditions imposed by the parent GCM, are very sensitive to some key parameters and require very advanced computing resources. The main strengths of statistical downscaling are that they account for the local scale of climate variability and are also computationally cheap. This low cost means that it is easy to create an ensemble of climate scenarios. The weaknesses are that they require long time-series of high quality observations, their performance depends on the choice of predictor variables and their statistical limits depend on the number of observations. Also, scenarios based on statistical downscaling methods only include future changes in statistical relationships between climate variables that are given as input to the method, whereas those from the physically based RCMs can simulate changes in such relationships a priori.

Until recently, most climate impact assessments were based on projections from GCMs. These projections are particularly inappropriate for lakes which are highly dynamic and respond in complex ways to short-term changes in weather. Therefore, downscaling of GCM projections has been an essential task in CLIME. In fact, CLIME was one of the first projects to take advantage of the new, high-resolution RCMs that are now available for the European region. Also the catchments modelled in CLIME typically responded in a non-linear way to changes in weather. Since lakes are very sensitive to short-term changes in the weather it is essential to model their response with as many combinations of daily weather sequences as possible. To meet the demands of these models, the RCM results were further downscaled by statistical methods using climate observations wherever there were long-term records of a high quality. The methods used varied from site to site and were largely dictated by the area covered and the length of the historical records (see Chapters 3, 9, 11 and 13 in this volume for further details).

2.4 The Downscaling Chain in CLIME

Figure 2.2 shows how the GCM outputs were downscaled to provide the high resolution time-series required by the lake and catchment modellers. At all sites, the primary variables used were the RCM outputs. At sites that covered a disproportionately large part of the region the raw RCM outputs were used to directly perturb the catchment models. At all other sites, the RCM outputs were statistically downscaled using an existing Weather Generator (WG) that was adapted to meet the needs of the CLIME project. A WG is a stochastic model that can be used to produce multiple realizations of weather based on the statistical relationships between the meteorological observations acquired at a specific site over an extended period of time. In practice, very long site-specific records are required to calibrate these models. At most CLIME sites, we used at least 20 years of daily measurements but slightly shorter records had to be used at a few locations. In the following section, we describe these downscaling procedures in detail and explain how these results were used to drive both the lake and catchment models.

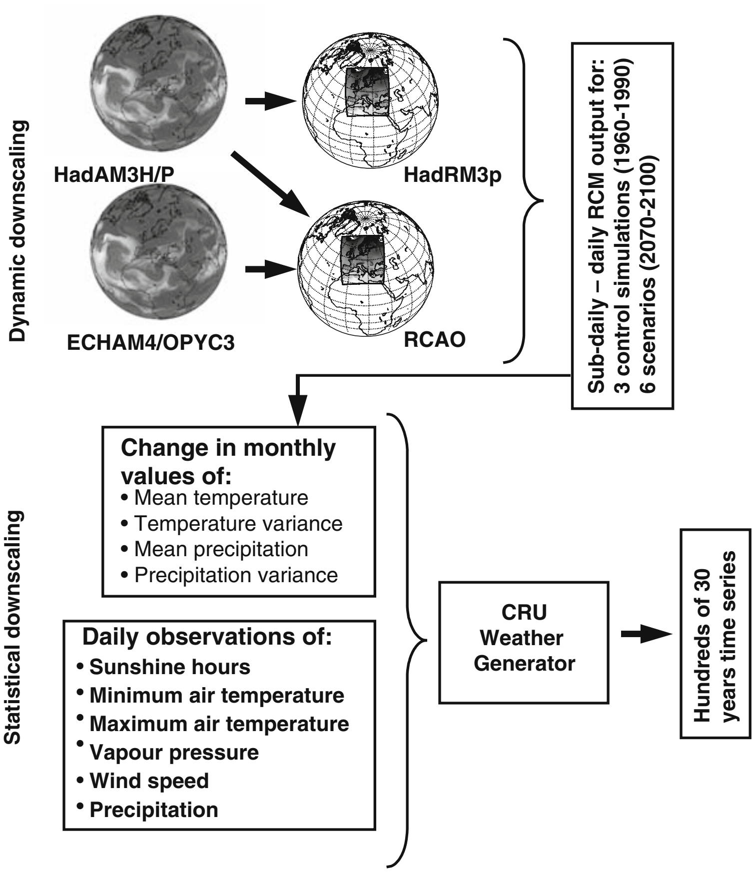

Fig. 2.2 A schematic diagram showing the downscaling methods used in CLIME. The parent GCMs were HadAM3H/P and ECHAM4/OPYC3. The RCMs used for dynamic downscaling were RCAO and HadRM3P. The statistical downscaling was based on the output from a weather generator developed by the Climatic Research Unit (CRU) in the UK

2.4.1 GCM Boundary Data

A regional climate model needs continuous forcing on its boundaries from gridded observations or from another model. In CLIME, data from two GCMs, HadAM3H/P and ECHAM4/OPYC3, were used to drive the RCMs (see the top left corner of Fig. 2.2). HadAM3H/P is the atmospheric component of the Hadley Centre coupled atmosphere-ocean GCM (AOGCM) HadCM3 developed by the UK Meteorological Office (Gordon et al., 2000). ECHAM4/OPYC3 is an AOGCM produced by the Max Planck Institute in Germany (Roeckner et al., 1999).

HadAM3H/P has a horizontal resolution of 1.875° longitude × 1.25° latitude. The sea surface temperature (SST) and sea ice distributions for the HadAM3H/P simulations were derived from observations and earlier, lower resolution, HadCM3 simulations. HadAM3P (Jones et al., 2004a) is an updated version of HadAM3H (Hudson and Jones, 2002). The update, concerning the representation of clouds and precipitation, was undertaken to correct biases in the moisture fields that were only observed in parts of the globe that lie outside Europe. The difference between the two versions, H and P, is quite small according to Frei et al. (2006) who compared the precipitation statistics for a complex Alpine region.

The atmospheric component of ECHAM4/OPYC3, ECHAM4, was run at T42 spectral resolution (equivalent to a grid spacing of 2.8° longitude × 2.8° latitude). To obtain a reasonable control climate in ECHAM4/OPYC3 it was necessary to apply corrections for the energy fluxes between the atmosphere and the ocean.

Both the AOGCMs, HadCM3 and ECHAM4/OPYC3, were first run from 1860 to 1990 using a combination of observed and estimated changes in the atmospheric composition. From 1990 on, each of these AOGCM runs was continued as two separate simulations using IPCC SRES scenarios A2 and B2 for anthropogenic greenhouse gas and sulphur emissions (Nakićenović et al., 2000). The A2 scenario assumes large and continuously increasing emissions of the major anthropogenic greenhouse gases CO_2, CH_4 and N_2O. The B2 scenario also assumes an increase in the CO_2 and CH_4 emissions, but at a slower rate i.e. in the lower midrange of the SRES scenarios. However, sulphur emissions are also larger in A2 than in B2, which partly compensates for the climatic effects of the larger greenhouse gas emissions in the A2 scenario.

2.4.2 Dynamic Downscaling

The dynamic downscaling in CLIME was based on two different Regional Climate Models (RCMs), the Rossby Centre Regional Climate Model (RCAO) developed by the Swedish Meteorological and Hydrological Institute and the Hadley Centre Regional Climate Model (HadRM3P).

The main components of RCAO (Räisänen et al., 2003, 2004) are the atmospheric model RCA2 (Jones et al., 2004b), the Baltic Sea ocean model RCO (Meier et al., 2003) and the lake model PROBE (Ljungemyr et al., 1996). RCA2 was run in a rotated longitude-latitude grid with 0.44° (ca. 49 km) resolution in both horizontal directions and with 24 levels in the vertical. The RCO simulations used covered the Baltic Sea. Sea-surface temperature outside the Baltic Sea was given by the GCM.

HadRM3P (Jones et al., 2004a) is the regional climate model developed by the UK Meteorological Office's Hadley Centre. Similar to RCA2, it is operated on a rotated longitude-latitude grid with 0.44° resolution in both horizontal directions and 19 levels in the vertical. Its dynamics and physical parameterizations are similar to HadAM3P and the sea-surface temperatures are those used in HadAM3P.

Table 2.1 lists all the climate model simulations used in CLIME. As illustrated at the top of Fig. 2.2 these were based on combinations of the two RCMs and the two GCMs including the two SRES emission scenarios. This produced a total of

Table 2.1 The RCM simulations used in CLIME

RCM	Simulation	Applied GCM	SRES Scenario	Abbreviation
RCAO	Control (1961–1990)	HadAM3H	–	RCAO-H-C
		ECHAM4/OPYC3	–	RCAO-E-C
	Scenario (2071–2100)	HadAM3H	A2	RCAO-H-A2
			B2	RCAO-H-B2
		ECHAM4/OPYC3	A2	RCAO-E-A2
			B2	RCAO-E-B2
HadRM3P	Control (1960–1989)	HadAM3P	–	HadRM3P-C
	Scenario (2070–2099)	HadAM3P	A2	HadRM3P-A2
			B2	HadRM3P-B2

three control simulations and six scenarios for the projected change in the climate. The control simulations cover the period 1960–1990 and the future scenarios the period 2070–2100. The RCMs use equivalent concentrations of greenhouse gases and aerosols for the selected periods as the ‘parent’ GCMs. As seen from the table, the RCAO used boundaries from HadAM3H while the HadRM3P used boundaries from HadAM3P. The output generated by these simulations included six hourly averages, daily averages and monthly mean values. Details of how these averages were used to drive the lake and catchment models are given in Chapters 3, 9, 11, 13 and 15 of this volume.

2.4.3 Statistical Downscaling

To explore the potential effect of climate change on the seasonal dynamics of the lakes and catchments, it is important to cover as many combinations of dry and wet periods as possible within the different temperature ranges. Since the RCM outputs only provide a limited number of weather combinations, a stochastic weather generator was used to downscale the RCM and produce multiple realisations of the weather on a catchment scale. A weather generator represents one among many methods of statistical downscaling.

The CRU daily weather generator (CRUWG) was initially developed by Jones and Salmon (1995) at the Climate Research Unit in the UK and later modified by Watts et al. (2004). The WG uses time-series of past meteorological measurements at a given site to estimate the parameters which are used to calibrate a stochastic model that then generates multiple streams of daily weather variables (Watts et al., 2004). In the CRUWG, precipitation is used as the primary driving variable. A first-order Markov chain model generates the amount of precipitation on any day by sampling at random from a gamma distribution. The distribution is formed by grouping each half month of data together i.e. the first half of the month is days 1–15 and the second half days 16–28, 29, 30 or 31. In the grouping, the data are also categorised into four states depending on how dry and wet days follow on each other.

The secondary weather variables produced by the weather generator include minimum temperature (°C), maximum temperature (°C), vapour pressure (hPa), wind speed (ms^{-1}) and sunshine duration (hours). Mean temperature, diurnal temperature range, vapour pressure, wind speed and sunshine duration are derived through first order auto-regressive processes. Minimum and maximum temperatures are then calculated from the mean temperature and diurnal temperature range and the relative humidity (%) derived from vapour pressure. The potential evapotranspiration (mm day^{-1}) is also calculated with respect to grass using the Penman-Monteith method.

Very long time series of daily meteorological measurements are required to calibrate or train the weather generator for the selected sites. The record must be long enough to capture the full range of variability and enough samples must be available for each combination of half month and state to form well defined gamma distributions. For sites characterised by long dry periods, too few samples may be available which results in poorly defined distributions which then generate time-series where the extremes are not properly represented. The recommendation by Watts et al. (2004) is to use at least 20–30 years of continuous data for calibration.

Table 2.2 lists the seven CLIME sites for which the CRUWG was used to produce multiple streams of daily weather. Since some of the catchment modellers required estimates of the frequency of extremely dry and wet periods, the weather generator was used to generate a minimum of 300 sequences, where each sequence was 30 years long. These time series were generated for the observational period itself and for the future conditions projected by the six RCM scenarios. For the future projections, the WG used the RCM changes in mean values and variances of temperature and precipitation, respectively, for the grid boxes nearest to the selected sites. As can be seen from the table, the record for one site was less than the recommended 20–30 years and that for another site was on the limit. As already discussed, too short time series may cause unrealistic generations of output due to ill defined distributions. Unfortunately we occasionally experienced unrealistic precipitation estimates in some long time-series, e.g. for Lough Leane. Since we wanted to use the same downscaling procedures at all the sites we filtered the time-series at the sites where the precipitation output included some non-realistic values. For WG precipitation

Table 2.2 The CLIME sites where the CRU Weather Generator was used to produce multiple realizations of the daily weather. Results for the locations denoted by * have been filtered as described in the text

Place	Position	Period	No of years
Lough Leane, Ireland*	52.03°N 9.33°W	1968–1997	30
Lough Feeagh, Ireland	53.95°N 9.58°W	1966–1993	28
Poulaphuca Reservoir, Ireland	53.08°N 6.30°W	1973–2002	30
Keswick, UK*	54.60°N 3.10°W	1984–2002	19
Tartu, Estonia*	58.27°N 26.47°E	1971–2003	33
Moor House, UK*	54.70°N 2.38°W	1983–2002	20
Västerås, Sweden*	59.59°N 16.60°E	1973–1997	25

values (WG P) that exceeded the maximum observed precipitation (observed P_{max}) the output was corrected as:

$$(WG\ P) = \frac{(WG\ P)}{(WG\ P_{\max})}(observed\ P_{\max}) * 1.1$$

At most sites this filter removed less than 0.1% of the samples but at Keswick the proportion approached 0.5%. For two sites with incomplete records the CRUWG could not be applied and the delta-change method was used instead. In the delta-change method the differences between an RCM scenario output and the corresponding RCM control output for a specific variable is applied to an observational time series of the same variable. The perturbed time series is then used to drive the models used to project future conditions. See Chapters 11 and 13 for further details and some applications of the delta-change method.

2.5 Climate Projections

In this section, the main features of the RCM and Weather Generator results are summarized using monthly and seasonal mean values for selected parameters. The first section reviews these results on a continental scale whilst the second summarizes some key features of the regional patterns.

2.5.1 European Continental Scale

Figure 2.3 shows the projected changes in the winter and summer temperature and precipitation for the European region. The maps show the average difference between each of the six individual scenarios and the corresponding control simulation. This representation was designed to show those areas where the models produced the most consistent results and also highlight those where there is more uncertainty.

The overall pattern can be explained by the general warming of the atmosphere, its large-scale dynamics and local feedback mechanisms. A warmer atmosphere can hold more water vapour. This means that the horizontal transport of water vapour will increase irrespective of any change in the general circulation pattern (Christensen et al., 2007). The increased transport will strengthen already established regions of convergence/divergence, such as the convergence in Northern Europe and divergence in Southern Europe. These changes results in increased amounts of precipitation over Northern Europe, especially during the winter, and decreased amounts of precipitation in Southern Europe, especially in summer. There is also some indication of a change in the large-scale dynamics, such as a pole-ward expansion of the subtropical highs and a pole-ward displacement of the mid-latitude westerlies and associated storm tracks (Christensen et al., 2007). These changes in dynamics can, in turn, either amplify or reduce the described tendencies in precipitation.

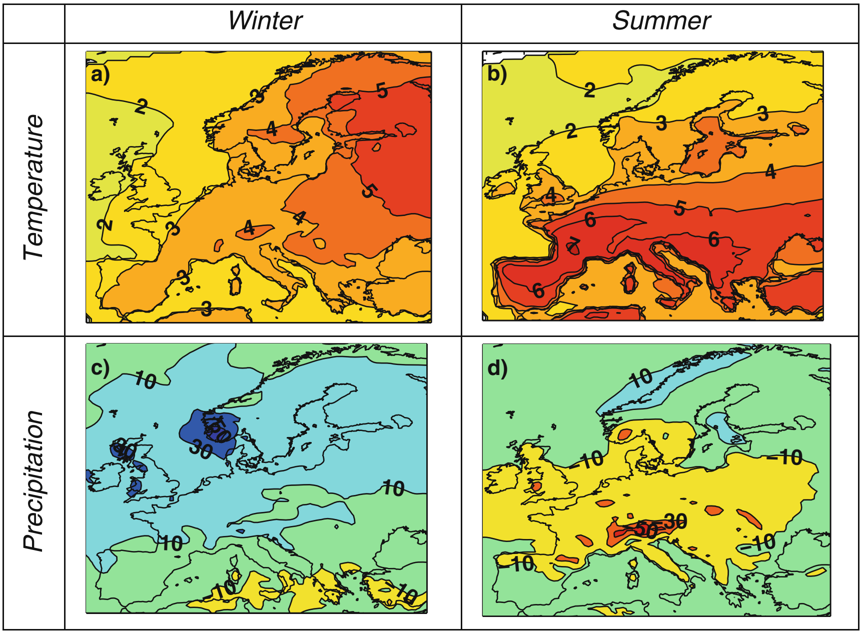

Fig. 2.3 The mean differences between the scenario (2071–2100) and control (1961–1990) outputs for all the simulations. The maps show the spatial differences in the winter (DJF) and summer (JJA) temperature (°C) and precipitation (mm/month)

The temperature variations show how the continental areas experience more pronounced warming than the maritime areas. This is due to the greater thermal inertia of the ocean as compared to land surfaces and that more energy is used for evaporation over the oceans than over the land. Apart from the land-sea contrast, the winter west-east warming gradient in Fig. 2.3a is mainly due to a shift of the snow line while the summer north–south gradient in Fig. 2.3b is mainly due to a drying out of Southern Europe (Déqué et al., 2005).

2.5.2 Clime Regional Scale

The lakes studied in CLIME were located in three very different climatic regions: the UK and Ireland, Northern Europe and Central Europe. In the following section, the RCM results will be presented, not individually for each lake, but as mean values for the grid boxes that cover these regions. Figure 2.4 shows the annual cycle, based on monthly means, of the projected change in a number of climatic parameters. For each sub figure, the results have been grouped according to the SRES emission scenario applied. The shaded areas thus cover the total range of values generated by all the RCMs for each SRES scenario (A2 or B2). Note that the area averaging of the parameters in each region is based on all the grid boxes that intersect with the drainage basin of the selected lakes.

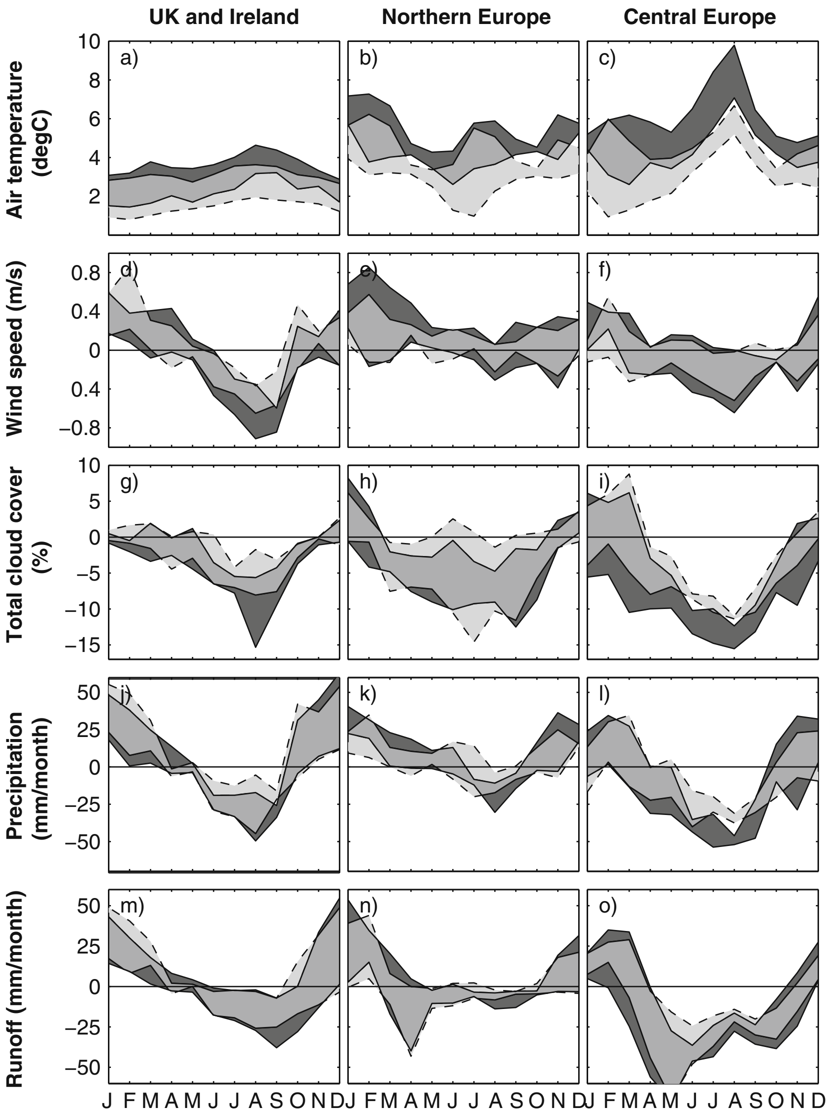

Fig. 2.4 The differences between the projected (2071–2100) and control (1961–1990) simulations for ensembles based on the A2 (*dark grey*) and B2 (*light grey*) emission scenarios for the regions that include CLIME sites. Note that the two distributions intersect in the *medium grey* areas. The results for each region are averaged over all grid squares containing a CLIME lake and drainage basin

In the UK and Ireland, the sites covered were Lough Feeagh, Lough Leane, Poulaphuca Reservoir, Esthwaite Water, Windermere, Bassenthwaite and Blelham Tarn. In Northern Europe, some of the lakes were much larger and the sites covered were Erken, Mälaren/Ekoln, Mälaren/Galten, Peipsi, Võrtsjärv, Pääjärvi, Ormajärvi and Valkea Kotinen. In Central Europe the lakes covered were Hechtsee, Lake Konstanz, Mondsee, Piburger See and Zürichsee in the Alpine region and Balaton in Hungary, a large but very shallow lake. Note, that the averages that appear in this figure for Central Europe are biased by the fact that most of the sites are located in the Alps, an area where very large reductions are projected for the summer rainfall. (see Fig. 2.3d).

2.5.2.1 Near-Surface Air Temperature

The most obvious effect of the changing climate is the sustained increase in the air temperature (Fig. 2.4a–c). The most pronounced increases are those projected for Central Europe during the summer. As expected, the A2 scenarios suggest more pronounced warming than the B2 scenarios. As concluded by Rowell and Jones (2006) the summer warming over Central Europe is larger than the global mean, a trend attributed to the projected drying out of the soil in the spring. In all three regions, the temperature extremes are projected to increase at a faster rate than the corresponding mean temperatures (Kjellström et al., 2007).

2.5.2.2 Wind Speed

The amplitude of the seasonal variation in the wind speed is projected to increase in all three regions (Fig 2.4d–f) the winters becoming windier and the summers calmer. The most pronounced difference is that projected for UK and Ireland where there is a marked reduction in wind speed during summer. Christensen et al. (2007) have, however, shown that these wind speed projections are very model dependent. Despite these uncertainties, there are indications of increased wind speed in Northern Europe and decreased wind speed in Southern Europe. Changes in extreme wind speeds tend to follow the changes in mean wind speeds in most of the models.

2.5.2.3 Cloudiness

All three regions show a reduction in summer cloudiness (Fig. 2.4g–i). The patterns projected for Northern Europe are quite variable but those for the Central European region show an obvious summer reduction, with small differences between the models and a large difference between the two emission scenarios. Such a reduction in summer cloudiness will of course lead to a corresponding increase in the solar radiation. The decrease in summer cloudiness over Central Europe is primarily a function of the spring-summer reduction in soil moisture.

2.5.2.4 Precipitation

Not surprisingly, the reduction in summer cloudiness is associated with a corresponding reduction in summer precipitation, a feature that is particularly pronounced over the Central Europe. During the winter, most of Europe will experience an increase in precipitation, especially in the north–west and in the north. According to Frei et al. (2006) the summer reductions in the mean precipitation projected for Central Europe can be explained by a decreased number of wet days while the increase of mean precipitation in the north can be explained by an increased number of wet days and an increase in the intensity of precipitation.

Extreme precipitation events show an increasing frequency all over Europe at all times of the year although this tendency is less robust than the changes in mean precipitation.

2.5.2.5 Runoff

In our scenarios, we defined runoff as the difference between precipitation and evapotranspiration, also taking into account any storage of water on the land surface and in the soil. Thus the projected increases in the winter precipitation will lead to particularly pronounced increase in runoff. In the colder parts of Europe, an increasing proportion of the winter precipitation will also fall as rain since the air temperatures will be significantly higher. This change from snow to rain will have a major effect on the local hydrology and could lead to an increased risk of flooding in some areas. The magnitude of the snow-melt peaks in Northern Europe and in the Alps will also be reduced and there will be a marked shift in the timing as winters become warmer and the spring temperatures increase.

2.5.3 Weather Generator Projections

Results based on the output from the CRU Weather Generator will not be discussed in this chapter. Further details can be found in Chapters 9, 11 and 13 which describe the application of the technique in the catchment models. The example in Fig. 2.5 compares the distribution functions generated by the CRUWG for the air temperature at Lough Feeagh with that recorded for the historical period. The CRUWG output is based on five synthesized sequences, each 30 years long. Changes in mean

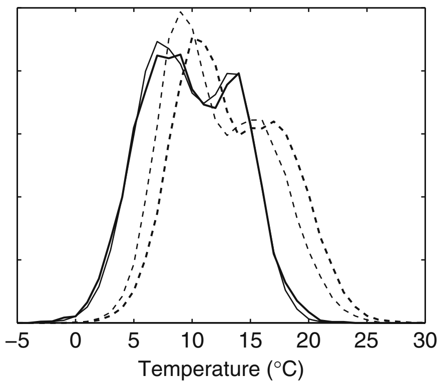

Fig. 2.5 Distribution functions of air temperature at Lough Feeagh, Ireland. The *solid lines* represent the control period for observations (*thick*) and WG output (*thin*). The *dashed lines* represent perturbed WG output based on the two scenarios HadRM3P-B2 (*thin*) and RCAO-E-A2 (*thick*)

values and variances of the air temperature, given from the RCM simulations, have been used for the perturbed distributions. Note that the form of each distribution and the magnitude of the shift towards the right changes systematically according to the chosen emission scenario.

2.6 Discussion and Concluding Remarks

The results presented here are based on a rather limited number of dynamical and statistical downscalings compared to the number currently available. The number of RCM downscalings has grown rapidly in the last few years. The outputs presented here nevertheless encapsulate a reasonable proportion of this variability and are broadly consistent with the patterns suggested in the fourth assessment report by IPCC (Christensen et al., 2007).

RCM outputs of this kind are now readily available via a number of web interfaces (e.g. that provided by the European PRUDENCE project). Recently, a number of transient scenarios have also been produced, i.e. scenarios where the simulations are not time slices but extend over a continuous period in time (e.g. 1960–2100). These outputs arrived too late for most of the model simulations undertaken in CLIME but an example of a simulation driven by one of these scenarios has been included in Chapter 15. In that example, a transient scenario was used to predict the long-term change in the concentration of oxygen in the deep water of a Swedish lake. The simulations showed a progressive reduction in the oxygen concentrations and suggested that very little oxygen would be present at the selected depth by 2020. If a 'time slice' model had been used for the same simulation it would simply have indicated zero values for the 2070–2100. Such results highlight the danger of assuming that natural systems respond in a linear way to the change in the climate.

In future, transient RCM scenarios will be available via the web site of the ENSEMBLES project (see References) and should prove very valuable for the next generation of climate impact studies. One study that has already accessed these results is Euro-limpacs (www.eurolimpacs.ucl.ac.uk), a project designed to explore the combined effects of climate change and other stressors on wetland systems as well as rivers and lakes. It is, however, important to emphasise that these outputs still have to be treated with some caution. Like all models simulating complex processes, RCMs have their strengths, weaknesses and inherent uncertainties. Our experience in CLIME demonstrates that it is essential for the limnological analysts and modellers to work closely with the climatologists in the project to make the best use of the RCM outputs and avoid any misunderstanding of the documented results. Particular attention also has to be paid to the uncertainties associated with the methods used for dynamic and statistical downscaling.

The degree of uncertainty associated with the dynamic downscaling of the GCM outputs with RCMs has been assessed by Déqué et al. (2007). They divided these areas of uncertainty into four groups: the choice of the GCM, the structure of the

RCM, the emission scenario and the initial conditions used in the RCM. Their analyses formed part of the PRUDENCE project and were based on 25 different RCM scenarios for the European region. They concluded that (i) the greatest source of uncertainty was the choice of GCM, (ii) the choice of RCM had less of an effect on the overall variability but had as large an effect on summer precipitation as the choice of GCM, (iii) the choice of emission scenario produced the largest uncertainty for summer temperature in southern Europe and (iv) the uncertainty due to the initial conditions was generally small compared with the other uncertainties. The scenarios used in CLIME were a sub-set of these 25 projections and we maximised the variance by combining the results from two different GCMs and two contrasting emission scenarios. The experience from PRUDENCE suggests that, whilst we can have high confidence in the ability of the models to simulate the seasonal variation in the mean temperature, the precipitation results are much more variable. Nevertheless the changes projected on a continental scale are more consistent and include the persistent drying in southern Europe and the general wetting of northern Europe (Déqué et al., 2005).

The results generated by statistical downscaling are also subject to considerable uncertainties. These uncertainties were explored in the EU-funded project STARDEX (see e.g. Haylock et al., 2006; Schmidli et al., 2007). The STARDEX project concluded that uncertainties can arise from: (i) the choice of statistical method, (ii) the choice of the predictor variables used to relate the local climate to the climate forcing and (iii) the scale the predictor variables represent in terms of climate model resolution and physical processes. In many cases, the inter-model differences in the downscaled outputs for a single station were just as large as those observed between different emissions scenarios. In many studies there is no single statistical downscaling method that can deal with all the spatial and temporal variation given in the climate regime. The ideal solution is to compare the results from a range of downscaling methods before selecting the one that best meets the needs of the end-user. Practical advice on the best methods to use can be found in two reports archived at the IPCC Data Distribution Centre. The first report (Mearns et al., 2003) includes useful guidelines on the application of dynamic downscaling whilst the second (Wilby et al., 2004) discusses the alternative statistical methods.

It is now clear that coupled numerical simulations run over several decades are required to properly represent the variations in current and future climates. The use of coupled models is essential because there are complex feed-back mechanisms taking place which affect the response of the system as a whole. Today, the term Earth System Models is used to reflect the status of the complexity reached. The definition of Earth System Models is not very precise but the name implies that these model systems not only account for the human impact on the environment but can also take account of the way human societies respond to environmental change. Although, these models include many processes they still operate on a relatively coarse scale so there is still a need to develop better ways of downscaling these outputs to meet the needs of environmental scientists and managers.

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Chapter 3
Modeling the Effects of Climate Change on Catchment Hydrology with the GWLF Model

Elliot Schneiderman, Marko Järvinen, Eleanor Jennings, Linda May, Karen Moore, Pamela S. Naden, and Don Pierson

3.1 Introduction

The influence of catchment hydrology on the volume and timing of water inputs to waterbodies, and on the material loads of nutrients, sediment, and pollutants is central to any assessment of the impact of climate change on lakes. Changes in the timing and amount of precipitation, particularly when coupled with a change in air temperature, influence all the major components of the hydrological cycle, including evapotranspiration, snow dynamics, soil moisture, groundwater storage, baseflow, surface runoff, and streamflow.

Most published studies on the effects of climate change on catchment-scale hydrology in Europe (Andreasson et al., 2004; Arnell, 1999; Bultot et al., 1992; Drogue et al., 2004; Eckhardt and Ulbrich, 2003; Gellens and Roulin, 1998; Graham, 2004; Charlton et al., 2006), Muller-Wohlfeil et al., 2000; Wilby and Harris, 2006) and elsewhere (Chiew et al., 1995; Dibike and Coulibaly, 2004; Lettenmaier et al., 1996; Loukas et al., 2002; Rosenberg et al,. 2003) utilise a common three-step approach to the analysis of the projected impacts. In the first step, a catchment hydrology simulation model is calibrated and tested against historical data. In the second step, a range of different climate models are run to generate future climate scenarios. In the third step, the catchment model is then run for both a control period and a future climate scenario to assess the potential impact of these changes on the hydrological output. This general approach was the one used in CLIME to quantify the effects of climate change on catchments located in the Western and Northern regions of Europe.

The catchment model used at all the sites was GWLF (Generalised Watershed Loading Functions Model) (Haith and Shoemaker, 1987; Haith et al., 1992; Schneiderman et al., 2002). GWLF is a lumped-parameter model that simulates a daily water balance and utilizes empirical loading functions to estimate the nutrient and

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sediment loads from a catchment. GWLF was chosen as a parsimonious yet robust model that could be adapted, calibrated, and applied to a variety of catchments situated in regions with very different climates. In this chapter we describe the hydrological component of the GWLF model and its implementation. Chapters 9, 11, and 13 describe the applications of GWLF for simulating the flux of dissolved inorganic phosphorus, dissolved inorganic nitrogen, and dissolved organic carbon respectively; applications that are all based on the hydrological routines described here.

In CLIME, the hydrologic effects of the projected changes in the climate were explored in four different catchments – two located in Western Europe and two in Northern Europe. The two Western catchments were the River Flesk, a sub-catchment of Lough Leane in SW Ireland (IE) and the Esthwaite catchment in the NW of the United Kingdom (UK). The two Northern catchments were the Arbogaån catchment of L. Mälaren in Sweden (SE) and the Mustajoki catchment of L. Pääjärvi in Finland (FI). Table 3.1 summarizes the key hydrological features of the four sites. The annual streamflow is from stream gauge observations, evapotranspiration is estimated by the temperature index method in the GWLF model, snow water equivalent is estimated from the GWLF simulations, and precipitation is based on meteorological data which is corrected during the model calibration (see later). Sites in the Western region are subject to more than twice the precipitation of that in the northern sites. Air temperatures are warmer at the Western sites, particularly in the winter. Streamflows per unit catchment area in the Western sites are more than three times greater than those in the Northern sites. There are also regional differences in the seasonality of precipitation, which influences the regional patterns in streamflow. Both regions have similar summer precipitation, but in the Northern sites summer precipitation is the greatest of any season whereas in the Western sites summer is the season with the least precipitation. Winter, spring and autumn pre-

Table 3.1 The hydrological characteristics of the selected catchments studied in Northern and Western regions of Europe

Country	Catchment	Position	Area (km)	Annual total precipitation (cm)	Annual snow water equivalent (cm)	Annual evapotranspiration (cm)	Annual streamflow (cm)
Western region							
Ireland	Flesk	52.03°N 9.33°W	325	210	0	58	152
UK	Esthwaite	54.60°N 3.10°W	15.6	221	4	45	176
Northern region							
Sweden	Arbogaån	59.59°N 16.6°E	3,808	90	21	48	42
Finland	Mustajoki	61.04°N 25.08°E	76.8	72	21	43	29

cipitation in the Western sites are twice those of the Northern sites. Under present climate conditions a significant portion of the winter precipitation in the Northern region falls as snow. This accumulates in the catchment until the spring when snowmelt augments streamflow to form a distinct snowmelt peak that often results in a maximum yearly streamflow.

3.2 The GWLF Catchment Model

GWLF is a conceptual lumped-parameter model (Leavesley, 1999) that simulates a daily water balance and partitions water among the different pathways of the hydrological cycle. The catchment is viewed as a system of different land areas (Hydrologic Response Units or HRUs) that produce direct runoff, and two lumped (i.e. averaged over the whole catchment) subsurface reservoirs: an unsaturated soil zone from which water can be lost by evapotranspiration (ET) and a deeper saturated zone that maintains the baseflow (Fig. 3.1). Concentrations of dissolved and suspended substances (e.g. nutrients and sediment) in streamflow are estimated at the catchment outlet by loading functions that empirically relate the concentrations in the two streamflow components (direct runoff and baseflow) to catchment and HRU-specific characteristics. This chapter describes the hydrological component of the GWLF model. Other chapters on nitrogen and phosphorus describe the loading function approach used by GWLF and its application in greater detail.

Different versions of the model, in addition to the original Haith model, have been developed and have been used for a variety of applications (Dai et al., 2000; Evans et al., 2002; Lee et al., 2000; Limbrunner et al., 2005a; Schneiderman et al., 2002; Smedberg et al., 2006). Each of these versions has particular modifications, but all follow the same GWLF conceptual framework. The CLIME project used the version developed by New York City Department of Environmental Protection for the NYC catchments (Schneiderman et al., 2002). This version has been used to evaluate the effects of land use, watershed management, and climate on water quality in the NYC water supply system. (NYC DEP, 2006).

The GWLF hydrology model is driven by daily time series of precipitation and air temperature that have been spatially averaged over the catchment. Precipitation occurs as rain or snow depending on daily air temperature. Snow water is stored in a lumped snowpack from which snowmelt is released as a function of air temperature.

Direct runoff for each HRU is estimated as a non-linear function of precipitation (P) and a catchment retention parameter S using the SCS Runoff Equation (USDA-SCS, 1972): *Direct Runoff* = $(P - 0.2^*S)^2/(\mathrm{P} + 0.8^*S)$ where P represents total water inputs, including rain and snowmelt. For a given HRU, S can vary between a minimum limit that applies when the catchment is considered wet and a maximum limit that applies when the catchment is dry. The catchment moisture condition is based on average 5-day antecedent. Catchment retention parameter values for average, wet, and dry catchment conditions are derived from runoff curve numbers, which have already been compiled for HRUs with different land use and soil

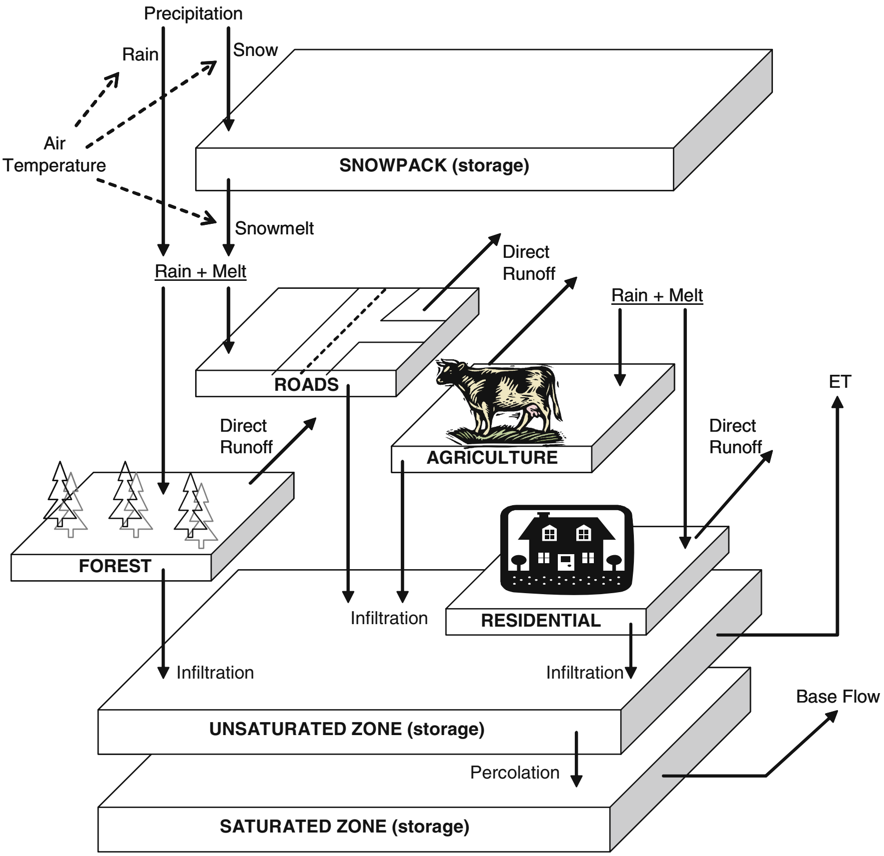

Fig. 3.1 Catchment hydrology as conceptualized in the GWLF Model (adapted from Limbrunner et al., 2005b)

type (USDA-SCS, 1986). For the CLIME simulations, the model was adapted to use the European Union's CORINE land use classification system (EEA, 2000). Curve numbers were derived from the literature and then scaled by site-specific calibration. In this scheme, direct runoff generated within the catchment is delayed by introducing a storage term to account for travel times that exceed the daily time step.

Rain and snowmelt that does not become direct runoff infiltrates into a lumped unsaturated soil zone storage, from which water is lost by evapotranspiration (*ET*). Potential evapotranspiration (*PET*) is calculated as a function of air temperature and daylength, using the Hamon temperature index equation (Hamon, 1961). The relationship between *ET* and *PET* depends upon the season, the landuse/landcover type and the available soil moisture. For each HRU, a seasonally-varying vegetative cover coefficient adjusts the value of *ET* for different land covers. A moisture extraction function (Shuttleworth, 1993) further reduces *ET* as soil moisture approaches wilting point and the availability of soil water to plants declines sharply. Water in the unsaturated zone that exceeds the average field capacity for soils within the catchment percolates to the lumped saturated zone storage. The baseflow component of streamflow is extracted from the saturated zone as a fraction (the recession coefficient) of the amount of water stored in that zone. Streamflow is calculated daily as the sum of baseflow and direct runoff.

Seven adjustable parameters are included in the GWLF hydrologic model. These parameters are: a coefficient for scaling the retention parameter S in the runoff equation, a snowmelt coefficient which determines rates of melt as a function of temperature, a runoff recession coefficient which describes the recession of direct runoff, two rate parameters (a recession and a slow recession coefficient), a baseflow storage capacity that control the baseflow output from the saturated zone and a precipitation correction factor to correct for any bias in the precipitation gauge.

For the CLIME simulations a GWLF model was set up, calibrated, and tested for each of the selected catchments. Continuous daily precipitation, air temperature, and stream flow data were gathered for an appropriate calibration and a separate validation period. Direct runoff and baseflow were estimated from daily streamflow data using the baseflow separation technique of Arnold et al. (1995). An iterative calibration of the model was then performed to minimise the mean square error between the simulated values and the measured variables. The precipitation correction factor was optimised in relation to the cumulated streamflow; the curve number adjustment in relation to the cumulated direct runoff; the runoff recession coefficient in relation to the daily direct runoff; the melt coefficient in relation to the daily streamflow; the recession coefficient in relation to the daily separated baseflow; and the slow recession coefficient and baseflow storage capacity in relation to the baseflow on the days when flow was less than a low flow threshold (defined as the 20th percentile of the daily flow values.) The resulting parameters are shown in Table 3.2 whilst Table 3.3 lists the curve numbers (scaled by the calibrated curve number adjustment) associated with the dominant land cover types in the four catchments.

The Nash-Sutcliffe model efficiency statistic E_{NS} (Nash and Sutcliffe, 1970) was used to measure goodness of fit, with a value of 1 signifying perfect fit and a value of zero or a negative value indicating that the mean observed data is a better pre-

Table 3.2 The catchment parameters used in the hydrological model

Catchment	Precip correct factor	Melt coeff cm/degC/ day	Runoff recess coeff/day	Soil water capacity* cm	Recess coeff/day	Slow recess coeff/day	Baseflow capacity cm
Flesk(IE)	1.09	0.25	0.73	15	0.116	0.005	20
Esthwaite (UK)	1.36	0.55	0.31	10	0.085	0.023	1.45
Arbogaån(SE)	1.14	0.22	0.10	15	0.017	0.002	5.56
Mustajoki(FI)	1.11	0.28	0.19	15	0.028	0.004	4.60

*set parameter; the rest are calibrated

Table 3.3 The dominant land cover types and scaled runoff curve numbers (CN) for the selected catchments. Arbogaån (SE) data is for Domsta subcatchment

	Esthwaite (UK)		Flesk (IE)		Mustajoki (FI)		Arbogaån (SE)	
Land use	% Area	CN	% Area	CN	% Area	CN	% Area	CN
Urban			2%	90.8			1%	77.3
Pasture	52%	87.2	31%	72.0			7%	76.0
Cultivated					13%	77.2		
Natural Grassland	10%	74.2	9%	72.0				
Forest	32%	81.2	6%	80.5	67%	82.2	78%	69.9
Moors/Heathland			17%	94.0				
Peat bog			29%	94.0	20%	93.2	6%	96.2
Water bodies	6%	98.0	1%	98.0			9%	98.0

dictor than the simulated results. The results of these tests (Table 3.4) indicate a reasonable model calibration and validation for all the selected catchments. The Nash-Sutcliffe efficiency values for streamflow ranged from 0.63 to 0.90 for the calibration period and 0.60 to 0.75 for the validation period. These are similar to those reported from other studies which have applied a variety of different models in comparable European catchments (e.g. Perrin et al., 2001; Parajka et al., 2005.) With regard to the partitioning of the flows, the degree of fit to the direct runoff is rather better in the western catchments than in the northern catchments which may relate to how well the models capture the timing and magnitude of snowmelt events. The Nash-Sutcliffe efficiencies for baseflow varied between 0.51 and 0.76 for the calibration period and 0.47 and 0.67 for the validation period. These values are again very similar to the results produced by other models (e.g. Rouhani et al., 2007).

Table 3.4 The calibration and validation results for the GWLF hydrology model

Catchment	Period	Streamflow E_{NS}	Direct Runoff E_{NS}	Baseflow E_{NS}
Flesk(IE)	Calib. 1996–2000	0.73	0.62	0.68
	Valid. 2002–2005	0.75	0.64	0.59
Esthwaite(UK)	Calib. 1992–1998	0.90	0.83	0.76
	Valid.	n/a	n/a	n/a
Arbogaån(SE)	Calib. 1981–1990	0.63	0.33	0.62
	Valid. 1991–2000	0.70	0.43	0.67
Mustajoki(FI)	Calib. 1982–2000	0.63	0.46	0.51
	Valid. 1993–2004	0.60	0.41	0.47

3.3 The Climate Change Scenarios

The six climate scenarios used in CLIME were based on two Global Climate Models (GCMs), two Regional Climate Models (RCMs) and two IPCC Emission Scenarios (Table 3.5). The GCMs used to provide the boundary conditions for the RCM's were the HadAM3H/P, from the Hadley Centre, UK, and ECHAM4/OPYC3, from

Table 3.5 The climate scenarios used in the CLIME project

Climate scenario	GCM	RCM	IPCC emission scenario
E A2	ECHAM4/OPYC3	RCAO	A2
E B2	ECHAM4/OPYC3	RCAO	B2
H A2	HadAM3h	RCAO	A2
H B2	HadAM3h	RCAO	B2
Had A2	HadAM3p	Had RM3p	A2
Had B2	HadAM3p	Had RM3p	B2

the Max Planck Institute, Germany. The RCMs used were the Rossby Centre RCM (RCAO) and the Hadley Centre RCM (HadRM3p). The IPCC Emission Scenarios were the A2 and B2 scenarios, defined as follows by the International Panel on Climate Change (IPCC) 'Special Report on Emission Scenarios' (http://www.ipcc.ch):

- A2 describes a heterogeneous world, with underlying theme of self-reliance and preservation of local identities. Global population rises continuously throughout the 21st century. The introduction of new technology is less rapid than in other scenarios.
- B2 describes a world with population also increasing throughout the 21st century but at a slower rate than A2.

Taken together, these six scenarios provide projections that encapsulate much of the variability and uncertainty associated with the current generation of climate simulations. A detailed account of these scenarios has already been given in Chapter 2 which also discusses the different sources of uncertainly associated with these projections. For each scenario, daily time series of the observed or projected variations in the local precipitation and air temperature were used to drive the catchment model. Each climate scenario was run for a time period that spanned a historical control period (1961–1990) and a future projection period (2071–2100).

Since the size of the grid boxes in the RCMs were generally much larger than the size of the catchments, two different methods were used to downscale the raw RCM outputs. At the two Western sites, we used a weather generator (WG) developed by the Climate Research Unit in the UK. At the two Northern sites, where insufficient historical data was available to train the WG, an alternative delta change method (Andréasson et al., 2004; Hay et al., 2000) was used to produce the required results. The weather generator was based on the version produced by Jones and Salmon (1995) and modified by Watts et al. (2004). The WG uses time-series of past meteorological measurements to estimate a series of statistical parameters which are then used to calibrate a stochastic model that generates multiple streams of daily weather variables (Watts et al., 2004). These outputs are the ones used to drive the 'control' simulations. The statistical distributions are subsequently modified to produce multiple realisations of the daily weather projected for the future climate (see Arnell, 2003). In the delta-change method the differences between the 'control' and 'future' outputs in the variables simulated by the RCMs is used to re-scale the observed time series to an observational time series for the same variables. Further details of the two methods are given in Chapter 2, this volume. Both methods effectively generate multiple (100) 30-year time series that have the same statistical characteristics (Prudhomme, 2006) but with a random component that represents the assumed natural variability.

In CLIME, the catchment models were run repeatedly using the multiple 30-year time-series that represented the variability associated with the different climatic scenarios. All land use related coefficients were maintained at the current values for all scenarios. The start and end days of the growing season for each combination were estimated using the method developed by Mitchell and Hulme (2002). Output statistics were compared on an annual and monthly basis to quantify the site-specific effects of the changing climate (Fig. 3.2).

3.4 The Effects of Climate Change on the Hydrological Regimes

The effects of climate change on the hydrological regimes in the four catchments were analyzed by examining annual and monthly water budgets for the control vs. the six climate change scenarios. Annual and monthly mean values were calculated for each of the 100 randomized 30 year model runs for each scenario and for each water budget component ($n = 3000$). Projected changes in median annual precipitation and streamflow varied widely depending on which GCM and RCM were used (Fig. 3.3). Scenarios based on the ECHAM4/OPYC3 GCM (E-A2/B2) consistently generated higher future estimates than those based on the HadAM3h/p GCMs (H-A2/B2, Had-A2/B2). These differences reflect the systematic differences in the atmospheric circulation generated by the two GCMs for the winter period (Räisänen et al., 2004)

The RCAO RCM scenarios (H-A2/B2) tended to produce higher estimates than those generated by the HadRM3p RCM (Had-A2/B2), particularly for the A2 emissions scenario. These differences highlight the necessity to use multiple RCM combinations to provide robust estimates when assessing the impacts of climate change on water budgets (Fowler et al., 2007).

A comparison of annual projections for A2 vs. B2 emission scenarios did not reveal a consistent pattern across all catchments. Therefore, to estimate potential future changes in annual and monthly water budgets, we compared the median value of the control with the range of the median values projected by all six climate scenarios. Figure 3.4 shows the range of projected changes in median annual water budget components for the six different climate change scenarios.

On an annual basis, all catchments showed projected increases in mean daily air temperature, with greater increases in Northern Europe. As PET is calculated in GWLF as a function of air temperature, the projected increases in PET and the

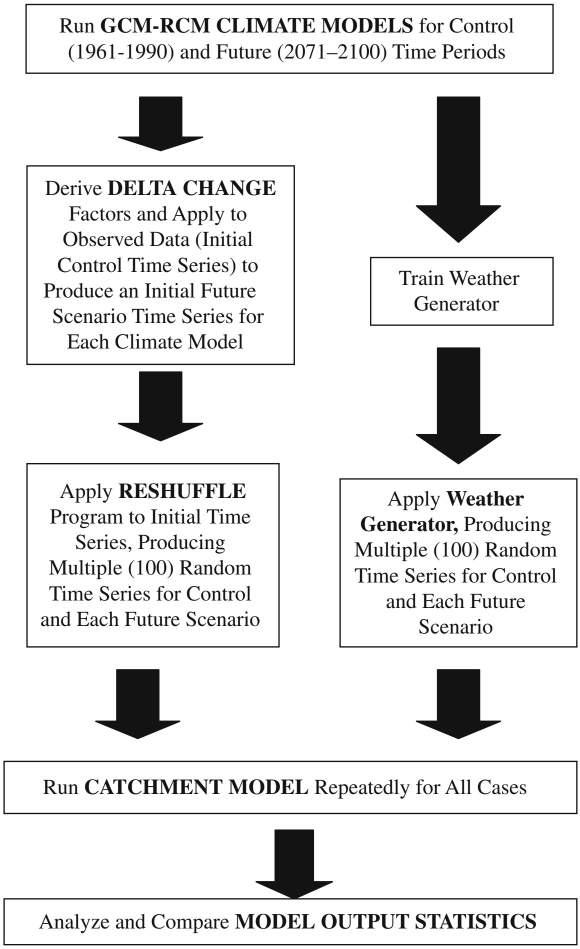

Fig. 3.2 The methods used to produced the climate change scenarios

associated decreases in soil water are a direct and consistent consequence of the increased air temperature. It should be noted however that these calculations do not take into account the impact of changes in net radiation, relative humidity or wind speed. ET projections differed from PET, as any reduction in the soil water content tends to reduce ET and thus offset the projected PET increases; this was particularly evident in the Flesk catchment (IE). It should also be noted that the model does not take into account any changes in land use or plant physiology that could occur in response to any change in the atmospheric concentration of CO_2 (Betts et al., 2007).

Projected change in annual precipitation was positive for Mustajoki (FI) but included negative as well as positive estimates for the other catchments. Streamflow projections followed those for annual precipitation. Projections for direct runoff and baseflow were more complex. Annual direct runoff increased in the Western

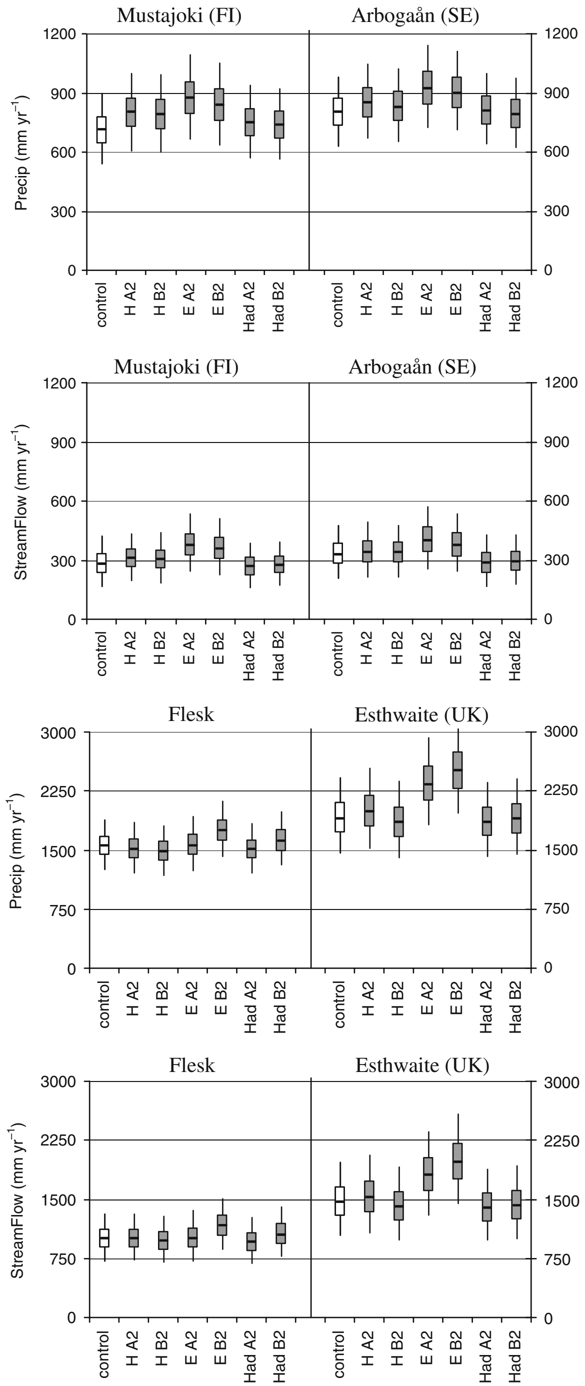

Fig. 3.3 The simulated variation in annual precipitation and streamflow (mm year-1) for control and six future climate scenarios for four study sites. Each *box* plot shows the variability of annual values for the one hundred 30-year ($n = 3000$) simulations with the 5–95 percentile range of variability (whiskers), the variability between the 25 and 75 percentile (*box*), and the median value. (Note difference in y-axis scale between northern and western catchments.)

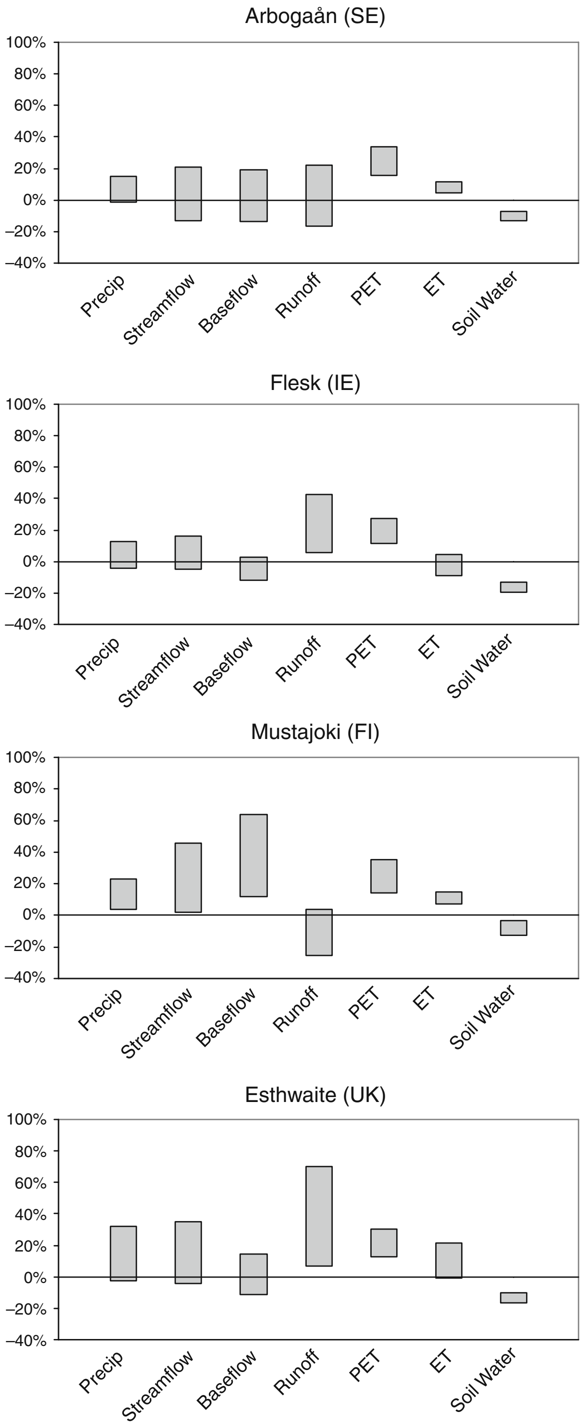

Fig. 3.4 Projected percent change in annual water budget components for four study sites. For each component, *vertical bar* shows the range of the percent change in the median values of these components for the six different scenarios of the future climate

catchments in response to increased dormant season precipitation, but in Mustajoki (FI) annual direct runoff decreased due to the sharp decline in the projected snowpack and the spring runoff event. Projected annual baseflow increased in Mustajoki (FI) due to increased winter infiltration (also related to change in snowpack), while annual baseflow projections for the other catchments were ambiguous.

The changes projected for the hydrologic regime reflected the projected seasonal variations in the air temperature and precipitation (Fig. 3.5a–b). While an increase in air temperature was projected throughout the year for all study catchments, greater increases were indicated in late summer for the two Western sites and in winter for the Northern sites with the most pronounced increases being suggested for Mustajoki (FI). Increased winter and decreased summer precipitation (Fig. 3.5b) was projected for all study catchments, with the seasonal differences being much more pronounced for the Western catchments. The increases in winter precipitation were greatest for Esthwaite (UK) (Fig. 3.5b). This site is located on the west coast of Britain, an area for which particularly high increases in winter rainfall are projected by all the scenarios described in Chapter 2.

Significant changes in both ET and soil moisture are projected for most of the study catchments. There was a pronounced increase in springtime ET for all catchments, followed by a decrease in summertime ET for all catchments except Mustajoki (FI) (Fig. 3.5d). The decline in summertime ET, in spite of increases in summertime PET (Fig. 3.5c), was largely controlled by the progressive depletion of the soil moisture during the growing season (Fig. 3.5e) and the associated decline in the transpiration rates as the soils dried out. The same pattern of depletion was indicated for all study catchments, but the decline was most pronounced at the Western sites, where the combined effects of increased summer air temperatures and decreased precipitation resulted in substantial declines in the soil moisture levels in summer and autumn.

In the Northern catchments, where snowmelt is a dominant hydrologic factor, the streamflow is strongly influenced by the dynamics of snow accumulation and melt. For these catchments, all six future climate scenarios projected large reductions in the size of the snowpack (Fig. 3.6d), with a reduction of more than 90% in maximum average snowpack in the months of January, February, and March. These snowpack reductions dominated the projected streamflow changes in all these catchments (Fig. 3.6d), with streamflow increasing during the winter and decreasing in spring. The baseflow (Fig. 3.6b) and runoff (Fig. 3.6c) components of streamflow both followed the same pattern. These effects were most pronounced in the Mustajoki (FI) catchment where the current streamflow is low in winter and increases rapidly during the spring snowmelt event.

In the Western catchments, where the hydrologic regimes are controlled by the rainfall, streamflow (Fig. 3.6a) is influenced by the seasonal interaction of precipitation and ET. Streamflow increased in winter and decreased in late summer and autumn, in accordance with changes in precipitation. However, the baseflow (Fig. 3.6b) and runoff (Fig. 3.6c) components of streamflow responded differently, with baseflow decreasing and runoff increasing in the fall and early winter, a combination that could result in increased streamflow flashiness (higher storm flows along

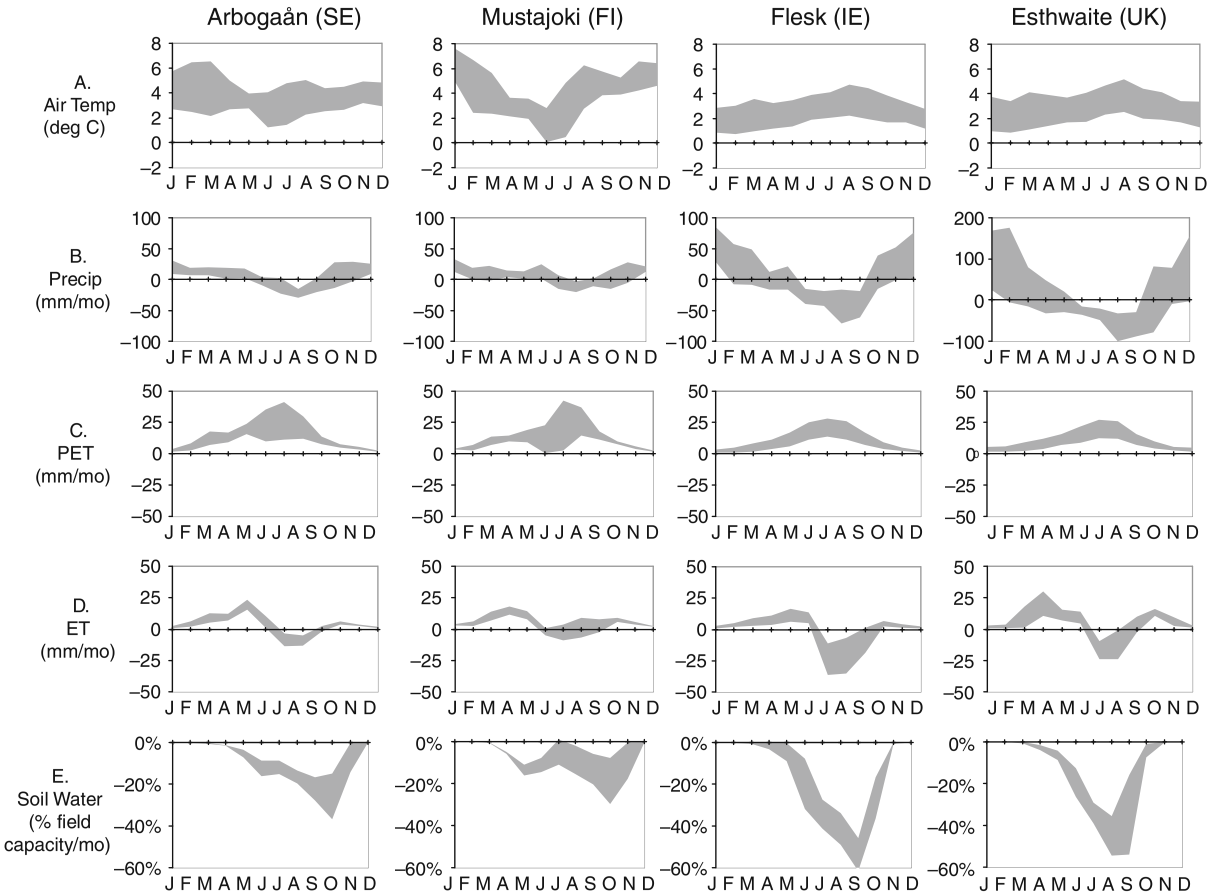

Fig. 3.5 Range of predicted changes of monthly air temperature (°C), precipitation (mm), PET (mm), ET (mm), and soil water (% field capacity) for the six climate change scenarios. Note differences in vertical scale for Esthwaite (UK)

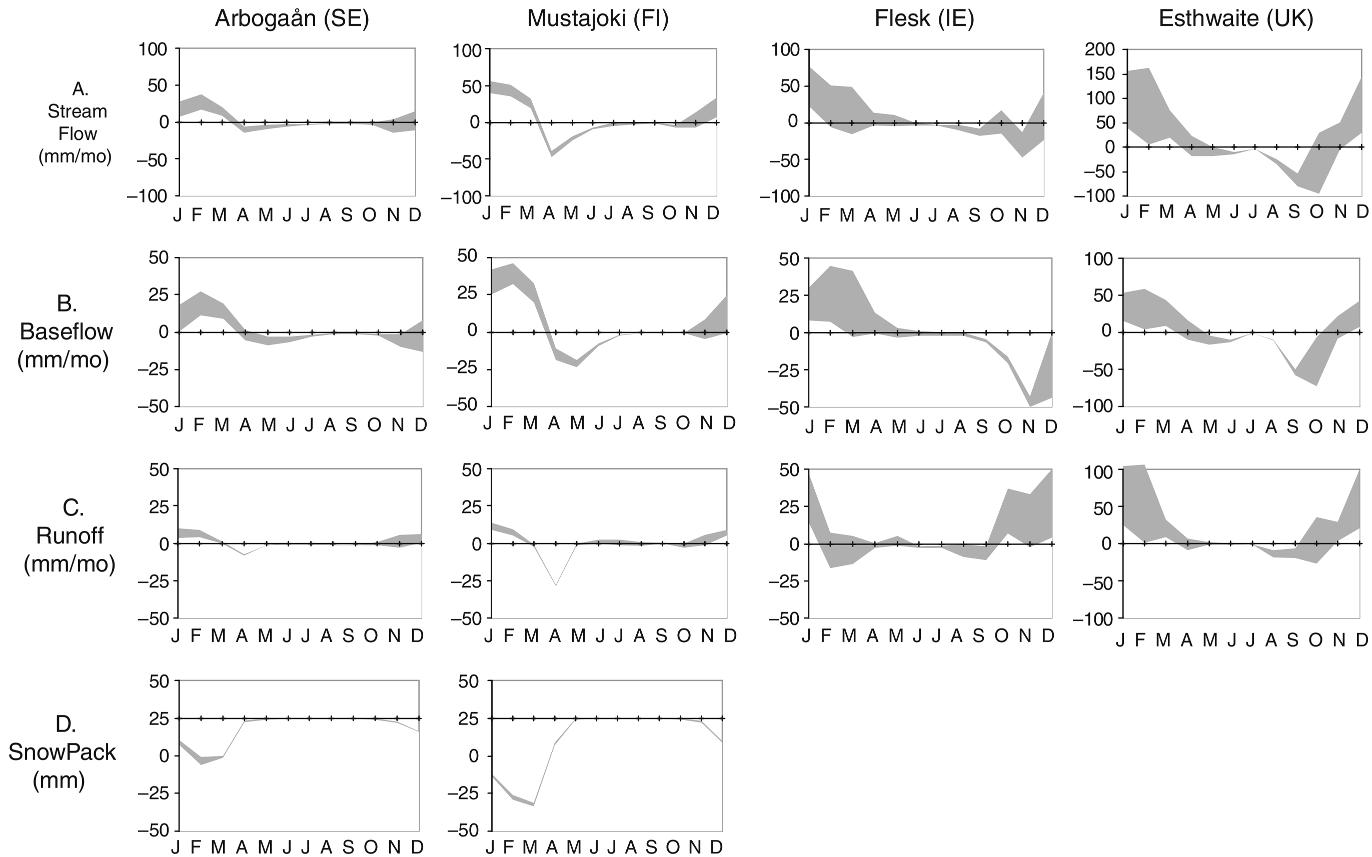

Fig. 3.6 Range of predicted changes of monthly streamflow (mm), baseflow (mm), runoff (mm), and snowpack (mm) for six climate change scenarios. Note differences in vertical scale for Esthwaite (UK)

with lower interstorm flows). During this period the runoff reflected the increase in precipitation, while baseflow declined due to the increased PET and the soil water depletion in summer and autumn.

3.5 Conclusions

Assessing the effects of climate change on catchment-scale hydrological regimes is complicated by the uncertainty of the future climate projections and the inherent complexity of hydrological systems. This complexity is highlighted by the differences in the projected streamflow generated in this study by the different combinations of RCMs and scenarios. The summary outputs presented here are based on multiple time series using four different climate models and two emission scenarios. Confidence in the seasonal and annual patterns revealed by these projections is greatly increased when the patterns reproduced by the different RCM-emission scenario combinations agree in direction. Conversely, our confidence in these projections tends to decrease when the climate models generate both increases and decreases for the same attribute. In general, however, the hydrological changes projected by these simulations are remarkably consistent and include clear directional shifts as well as some contradictory changes (Figs. 3.4, 3.5 and 3.6).

The effects of climate change on the hydrological characteristics of a catchment are mediated by the interaction of the meteorological forcing and the fluxes and storages in the catchment (Fig. 3.1). The projected responses vary of storages and fluxes to climate change from catchment to catchment and region to region. In Northern Europe, where snow is a dominant feature, rising winter air temperatures were projected to reduce the snowpack, increase the winter streamflow (baseflow and runoff) and lead to a reduction in the spring streamflow peak due to the marked change in seasonality. In Western Europe, the soil water storage plays a key role in mediating climate change effects.

In both the Northern and Western catchments, the quantity of water stored in the soil depends on the combined effects of precipitation and air temperature. The projected combination of higher summer air temperatures and reduced summer precipitation will, in future, result in substantial reductions in summer soil water storage. The severity of this summer soil water deficit and the time required to return to field capacity in autumn affects can also be expected to vary in different systems. For Western Europe the simulations presented here demonstrate that the recovery of the soil water may be delayed to such an extent that the baseflow in late autumn and early winter is reduced despite the increased precipitation. These increases in the flashiness of these systems could have important implications for future variations in water quality (Chapters 9 and 11), water colour (Chapter 13) and the seasonal development of phytoplankton (Chapter 1). In contrast, less change in future soil water deficit was projected for the Mustajoki (FI) catchment, and late autumn streamflow rose with increased precipitation but without increased flashiness. These examples highlight the need to consider the complexity of the interactions that exist between the local hydrology of and the regional climate forcing. It is also important to apply the selected models to as many different systems as possible to quantify their sensitivity to both long-term and short-term changes in the climate.

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Chapter 4
Lake Ice Phenology

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4.1 Introduction

In Chapter 5 of this book, it is shown that the formation of ice on the surface of a lake ('ice-on') and its thawing and ultimate disappearance ('ice-off') are complex phenomena governed by mechanisms that involve many interacting meteorological (and some non-meteorological) forcing factors. Linking ice phenology – the timing of ice-on and ice-off – to climatic forcing might therefore be expected to be a difficult task. This task, however, is simplified considerably by the fact that air temperature is the dominant variable driving ice phenology (Williams, 1971; Ruosteenoja, 1986; Vavrus et al., 1996; Williams and Stefan, 2006), and is also correlated to some extent with other relevant meteorological driving variables such as solar radiation, relative humidity and snowfall. Statistically, air temperature is often able to explain 60–70% of the variance in the timing of ice-off on lakes (Palecki and Barry, 1986; Livingstone, 1997). In a simplified sense, ice phenology can therefore often be viewed empirically as a temporally integrated response to the seasonal cycle in air temperature, modified by interannual differences in this cycle. Thus, despite the influence of geographical factors such as lake morphometry (Stewart and Haugen, 1990; Jensen et al., 2007) and meteorological factors such as snow cover (Vavrus et al., 1996; Jensen et al., 2007), it is often possible not only to model the timing of ice-on and ice-off empirically based on air temperature alone, but also to do the reverse; i.e., to use historical observations of ice phenology to draw conclusions about ambient air temperatures in the past (Magnuson et al., 2000b). This is particularly useful because air temperature exhibits a high degree of spatial coherence, so that historical variations in lake ice phenology reflect variations in air temperature on very large spatial scales (e.g., Livingstone, 1997, 1999, 2000; Magnuson et al., 2004, 2005; Jensen et al., 2007). Long time-series of ice phenology data have been employed successfully as indicators of past regional climate in Finland (Simojoki,

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1940; Palecki and Barry, 1986; Ruosteenoja, 1986; Kuusisto, 1987, 1993); Japan (Arakawa, 1954; Gray, 1974; Tanaka and Yoshino, 1982; Gordon et al., 1985); Switzerland (Pfister, 1984; Livingstone, 1997); various parts of Canada (Williams, 1971; Tramoni et al., 1985; Skinner, 1986, 1993); and the Great Lakes region of the USA (Robertson et al., 1992; Assel and Robertson, 1995).

4.2 Large-Scale Climatic Influences on Lake Ice Phenology

In the long term, the secular increase in global mean air temperature that has been occurring since the 19th century (Trenberth et al., 2007) appears to have affected the timing of both ice-on and ice-off. An analysis of long (150-year) time-series of observations of ice phenology from lakes and rivers throughout the Northern Hemisphere by Magnuson et al. (2000a) indicated that ice-on has been occurring later in the year at a mean rate of 5.7 days per 100 years, while ice-off has been occurring earlier in the year at a mean rate of 6.3 days per 100 years, implying an overall decrease in the duration of ice cover at a mean rate of 12 days per 100 years. If lake ice phenology responds to large-scale climate, then factors known to affect large-scale climate should also be detectable in ice phenology time-series. One such factor is global volcanism. Studies have shown that fluctuations in global volcanic activity are indeed reflected in fluctuations in the timing of ice-off on a high-altitude lake in Switzerland (Livingstone, 1997), and can also be detected, albeit less strongly, in other Northern Hemisphere lakes located at lower altitudes (Livingstone, 2000).

Several studies have demonstrated convincingly that the El Niño/Southern Oscillation (ENSO) phenomenon, which is known to have a strong influence on climate in the Americas and eastern Asia, leaves a pronounced signature in the ice phenology of many North American lakes (e.g., Anderson et al., 1996; Robertson et al., 2000; Bonsal et al., 2006), as does another large-scale climate phenomenon, the Pacific North America (PNA) pattern (Benson et al., 2000; Bonsal et al., 2006). In Europe and northern Asia, the North Atlantic Oscillation (NAO) (or the Arctic Oscillation, AO, with which it is strongly associated) is known to influence large-scale climate, especially air temperature, in winter and spring (e.g., Hurrell, 1995). It is thus likely that the winter NAO (or AO) will leave its signature in the timing of ice-off in high-latitude and high-altitude lakes in these areas. Various studies have confirmed this. An extremely strong winter NAO/AO signal has been detected in the ice phenology of lakes and rivers in Estonia, Finland and Sweden (Livingstone, 2000; Yoo and D'Odorico, 2002; Blenckner et al., 2004), and a strong NAO/AO signal is present in the ice phenology of Lake Baikal in Siberia (Livingstone, 1999; Todd and Mackay, 2003). Figure 4.1 illustrates two examples of the dependence of the duration of ice cover on the winter NAO index exhibited by a lake in Sweden (Lake Erken) and a lake in Germany (Müggelsee); in both cases, the dependence is approximately linear.

Recent work has shown that the influence of the winter NAO on ice phenology can be detected even in western European lakes that are situated in a mild, maritime

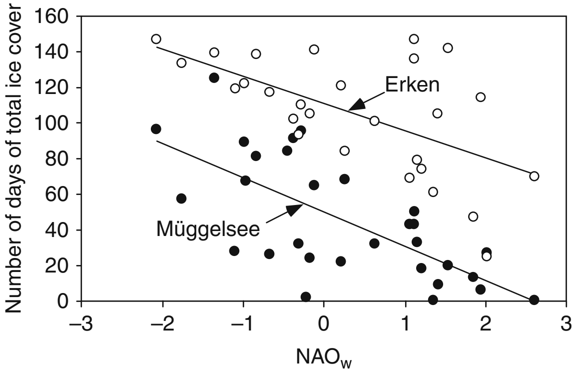
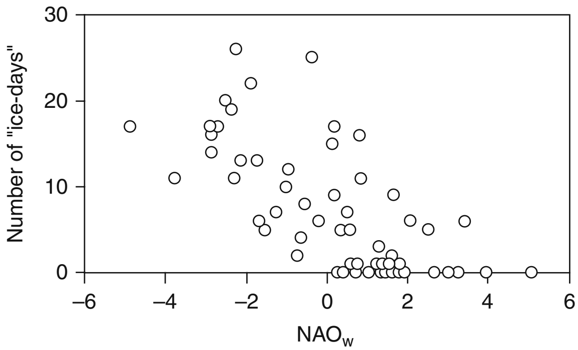


Fig. 4.1 Dependence of the number of days of total ice cover on Lake Erken (Sweden) and Müggelsee (northern Germany) on Hurrell's (1995) index of the North Atlantic Oscillation in winter (NAO_W), based on data from the winter of 1976/1977 to the winter of 2005/2006. The interannual variability of NAO_W explains 35% of the interannual variability of the duration of ice cover in the case of Lake Erken and 47% in the case of Müggelsee (the linear regression lines are illustrated; $p < 0.01$)

Fig. 4.2 Relationship between the number of 'ice-days' for Windermere (the number of days in winter on which the presence of any ice was recorded) and Hurrell's (1995) index of the North Atlantic Oscillation in winter (NAO_W). Based on data from 1933 to 2000. After George (2007)



climate and hence experience only short periods of ice cover, and then only in sheltered bays. George (2007) showed that the number of days per winter on which ice was recorded in two bays of Windermere, a lake in the English Lake District, was significantly related to regional air temperature and to the winter NAO (Fig. 4.2), and was also significantly correlated with the maximum annual extent of ice in the Baltic Sea, which is known to be influenced strongly by the winter NAO. The pronounced relationship to the NAO explained 50% of the interannual variance in the number of such 'ice-days' observed for Windermere.

A comparison of the timing of ice-out on Northern Hemisphere lakes since the middle of the 19th century with indices of the NAO in winter (Livingstone, 2000) has shown that the strength of this relationship has varied with time in some geographical regions. Although the relationship appears always to have been strong in northern European lakes, for lakes in Switzerland and Siberia the relationship is substantially stronger now than it was previously. By contrast, in the case of Lake Mendota, Wisconsin, the strength of the relationship of the timing of ice-out to winter NAO indices appears to have weakened considerably since the 19th century, while the relationship with the Southern Oscillation has simultaneously strengthened (Livingstone, 2000). It cannot therefore be assumed that the effect of the NAO on lake ice phenology in future will necessarily be the same as it is today, thus complicating any prediction of the impact of climate change on lake ice phenology.

4.3 Dependence of Sensitivity of the Timing of Ice-on and Ice-off on Air Temperature

Because air temperature varies sinusoidally during the year, the calendar dates on which the smoothed air temperature falls below 0°C (in winter) and rises above 0°C (in spring) are arc cosine functions of the smoothed air temperature, and the fraction of the year during which the ambient air temperature is below 0°C can be estimated by $(1/\pi)\ \arccos(T_m/T_a)$, where T_m and T_a are the mean and amplitude of the annual air temperature cycle. Assuming that the calendar dates of ice-on and ice-off vary linearly with the calendar dates on which the smoothed air temperature passes through 0°C (an assumption supported by the results of Duguay et al. (2006), who showed that the spatial patterns of trends in ice-on and ice-off dates in Canada closely follow those of the 0°C isotherm in autumn and spring, respectively), it follows that the timing of ice-on and ice-off can also be described well by arc cosine functions of T_m and T_a. Based on over 40 years of historical ice phenology data from 196 lakes spanning 13° of latitude in Sweden, Weyhenmeyer et al. (2004) confirmed that the relationship between the timing of ice-off and air temperature is indeed described well by an arc cosine function (Fig. 4.3). Because of the form of the arc cosine function, the timing of ice-off (and the timing of ice-on) responds more sensitively to variations in ambient air temperature at lower latitudes, where mean annual air temperatures are higher, than at higher latitudes, where mean annual air temperatures are lower. The timing of ice-off therefore responds much more strongly to interannual variations in air temperature in southern Sweden, where winters are relatively mild and the duration of ice cover usually varies between 0 and 125 days, than in northern Sweden, where winters are more severe and ice cover usually lasts from 200 to 250 days. Thus the sensitivity of the timing of ice-off to air temperature is ~14 days per 1°C in southern Sweden, but only ~4 days per 1°C in northern Sweden.

This has further implications, however, because as air temperatures rise as a result of climate change, the sensitivity of the duration of ice cover to air temperature fluctuations will also increase. Thus, as air temperatures rise, the timing of ice-on and ice-off will shift more rapidly on lakes which now experience short periods of ice cover than on lakes which now experience long periods of ice cover. Again turning to the example of Sweden, it would be expected that gradually increasing

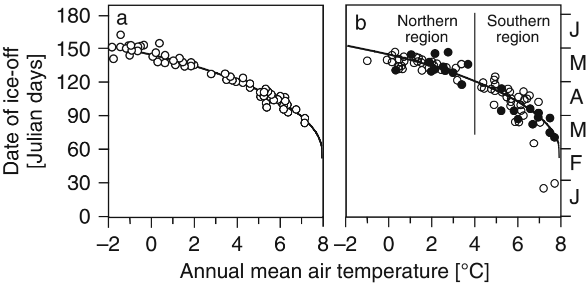


Fig. 4.3 (**a**) Spatial variation in the relationship between the median date of ice-off and lake-specific annual mean air temperature (1961–1990) for 70 lakes in Sweden, based on at least 28 years of ice phenology data for each lake. (**b**) Temporal variation in the relationship between the annual date of ice-off (median values for 14 lakes in a northern region and 14 lakes in a southern region of Sweden) and July-June annual mean air temperatures from 1961–1990 (*white*) and from 1991 to 2002 (*black*). The curves illustrated are based on the arc cosine model mentioned in the text. After Weyhenmeyer et al. (2004)

air temperatures will result in stronger shifts in the timing of ice-off on lakes in southern Sweden than northern Sweden. This was investigated by Weyhenmeyer et al. (2005), based on ice phenology observations from 54 Swedish lakes for the period 1961–1990. During this period, surface air temperatures in the Northern Hemisphere exhibited an increasing trend. This was reflected in significant trends towards earlier ice-off in 47 of the 54 lakes in the study. As expected, these trends were not uniform, but were much stronger in the southern lakes than the northern lakes. An increase in the mean air temperature of the Northern Hemisphere of 0.4 K during 1961–1990 was reflected in a shift in the timing of ice-off by up to ~70 days in southern Sweden, but only by ~10 days in northern Sweden. Further confirmation that the sensitivity of lake ice phenology to increasing air temperatures is greater at low latitudes than at higher latitudes was provided by Korhonen (2006), who showed that lakes in southern Finland have undergone a stronger long-term shift towards earlier ice break-up than lakes in the rest of the country. In North America, Hodgkins et al. (2002) found that the timing of ice-off on lakes in southern New England reponded more sensitively to air temperature than lakes in the northern, mountainous part of the region, and Jensen et al. (2007) showed for the Laurentian Great Lakes region that the substantial shift towards earlier ice-off that has been recorded over the past few decades is occurring most strongly at the southern boundary of the area in which lakes are routinely ice-covered during winter. Future increases in air temperature associated with climate change are therefore likely to result not only in generally shorter periods of ice cover on all lakes, but also in a disproportionately rapid decrease in the duration of ice cover on lakes that already have only a short ice season, potentially changing them from strictly dimictic, ice-covered lakes into monomictic, open-water lakes. Since the presence or absence of winter ice cover has wide-ranging ramifications for vertical mixing, deep-water oxygenation, nutrient recycling and algal productivity, this may lead to an alteration in the ecological status of ice-covered lakes in temperate regions. By contrast, increases in mean annual air temperature will have much less of an effect on lakes in very cold regions.

Modifying the arc cosine model to take account of the stochastic variability of air temperature around the annual sinusoid results in a vast improvement to the model, which allows the total duration of ice cover in any winter to be estimated in terms of an air temperature probability function (Livingstone and Adrian, 2009). Using this probability model it is possible to predict the total duration of ice cover on a lake even when the ice cover is intermittent; i.e., when periods of ice cover during winter are interrupted by periods of open water. This approach was tested successfully by Livingstone and Adrian (2009) on Müggelsee, a shallow lake in northern Germany which, depending on the severity of the winter, can experience either no ice cover, intermittent ice cover or continuous ice cover (Adrian and Hintze, 2000). Using regional climate model air temperature forecasts based on IPCC SRES climate scenarios A2 and B2, the probability model predicts that for Müggelsee, the percentage of ice-free winters will increase from ~2% now to over 60% by the end of the 21st century. Under current climate conditions, this would correspond to shifting the lake southwards by about 800 km to northern Italy.

4.4 Effects of Altitude on Lake Ice Phenology

Ice phenology depends not only on latitude, but also on altitude. Unsurprisingly, the duration of ice cover is found to increase with increasing altitude above sea-level (Eckel, 1955; Livingstone and Dokulil, 2001). A recent investigation into the timing of ice-on and ice-off on a suite of 19 high-altitude lakes in the Tatra Mountains during one ice season confirmed this, but also demonstrated that the altitudinal dependence was completely different for the timing of ice-on and the timing of ice-off (Šporka et al., 2006). The lakes spanned altitudes from 1,580 to 2,157 m a.s.l., and the duration of ice cover, which varied from 136 days to 232 days, exhibited a significant linear dependence on altitude, increasing by 10.2 days per 100 m (Fig. 4.4a). The timing of ice-off was also strongly dependent on altitude (Fig. 4.4b), with the highest lakes thawing much later than the lower lakes: e.g., lakes lying between 2,100 and 2,150 m a.s.l. thawed 47 days later than lakes lying between 1,550 and 1,700 m a.s.l. The overall rate of change of the timing of ice-off with altitude was 9.1 days per 100 m. However, although the timing of ice-on varied by up to 59 days among the lakes, it showed no detectable dependence on altitude (Fig. 4.4c). Šporka et al. (2006) explain the difference in behaviour of ice-on and ice-off as the consequence of a difference in the degree of their dependence on internal lake properties and processes. These are important determinants of freezing, but not of thawing, which is governed much more directly by external meteorological forcing acting to a large extent via altitudinally-dependent air temperature. (Seasonal differences in the frequency of occurrence of air temperature inversions, which are much more common in autumn and winter than in spring and summer, may also play a minor role here). Because the timing of ice-on is essentially independent of altitude, the altitudinal dependence of the duration of ice cover results solely from

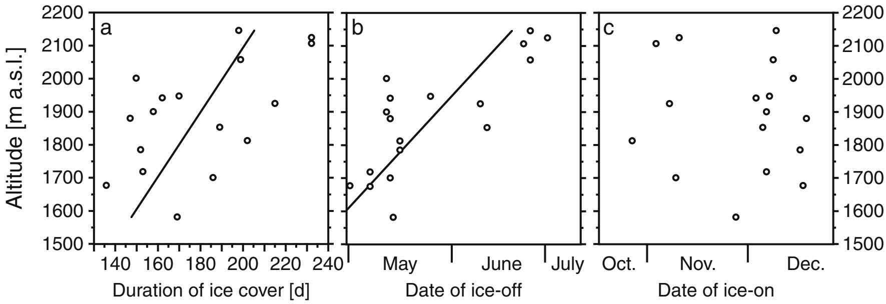


Fig. 4.4 Dependence of ice phenology on altitude for high-altitude lakes in the Tatra Mountains of Slovakia and Poland. (**a**) With increasing altitude, the duration of ice cover increases by 10.2 days per 100 m. (**b**) With increasing altitude, ice-off becomes later at the rate of 9.1 days per 100 m. (**c**) The timing of ice-on exhibits no significant dependence on altitude. After Šporka et al. (2006)

the altitudinal dependence of the timing of ice-off. Generalising from this, it can be hypothesized that variations in air temperature are much more likely to be reflected clearly in the timing of ice-off than in the timing of ice-on, implying that the former is probably a more direct, and hence more useful, climate indicator than the latter. In addition, it is likely that climate warming will affect the timing of ice-off in a much more predictable fashion than the timing of ice-on, which depends to a large extent on individual lake characteristics. This has implications for upscaling from the behaviour of one lake to the behaviour of a lake district: while the timing of ice-off is likely to vary coherently within a lake district, the timing of ice-on probably will not. This hypothesis is supported by the results of an analysis of recent trends in the ice phenology of lakes in Canada by Duguay et al. (2006), who found that ice-on dates generally exhibited lower regional coherence and showed fewer clear temporal trends than ice-off dates.

4.5 Conclusions

The importance of historical observations of lake ice phenology as a valuable store of information on past trends and fluctuations in large-scale climate is now well-established (e.g., Magnuson et al., 2000a, 2000b). Ice-on has been occurring progressively later and ice-off progressively earlier in various regions of the Northern Hemisphere since at least the middle of the 19th century (Magnuson et al., 2000a). The gradual shifts in ice phenology recorded during the 19th century may possibly represent a natural response to the termination of the Little Ice Age; subsequently, however, the rates at which the timing of ice-on and ice-off are shifting appear to have accelerated, with extremely high rates of change being reported in some regions of the Northern Hemisphere over the past few decades (e.g., Weyhenmeyer et al., 2005; Johnson and Stefan, 2006; Korhonen, 2006; Jensen et al., 2007). Given the weight of evidence supporting the hypothesis of anthropogenically forced global climate change (Trenberth et al., 2007), climate warming is likely to be the most important factor responsible for these recent shifts in lake ice phenology, although regional differences may reflect the differing importance of forcing by large-scale climate drivers such as ENSO and the NAO/AO.

Because the timing and duration of ice cover are important determinants of the ecology and water quality of high-latitude and high-altitude lakes (e.g., Adrian et al., 1999; Weyhenmeyer et al., 1999), it is important to be able to predict the future effects of climate change on lake ice phenology. Austin and Colman (2007) have shown for Lake Superior that surface water temperatures in summer are partially dependent on the duration and spatial extent of ice cover during the previous winter, so that diminishing ice cover is contributing to a long-term rise in summer water temperatures, and thus directly affecting the physical environment of the aquatic biota during a large part of the year. The recent results described in this chapter suggest that the impact of future climate change on lake ice phenology will not be uniform. It appears likely that climate warming will have its greatest effect on the timing and duration of ice cover on temperate lakes that already have a short ice season, whereas the timing and duration of ice cover on high-latitude or high-altitude lakes with a long ice season will be much less affected. The fact that the (spatial) increase in air temperature that occurs with decreasing altitude above sea level is reflected only in the timing of ice-off, but not in the timing of ice-on, suggests that the (temporal) increase in air temperature associated with climate change is likely to have a clearer impact on the timing of ice-off on lakes than on the timing of ice-on.

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Chapter 5
Modelling the Formation and Decay of Lake Ice

Matti Leppäranta

5.1 Introduction

In northern and mountainous regions of Europe lakes are frozen in winter. The ice season can be up to seven months long and the thickness of the ice can reach 100 cm. The physical characteristics of the water under the ice are very different to that found in the open water. The ice cover stabilizes the thermal characteristics of the lake, the surface water is kept at the freezing point and there is very little vertical transfer of heat. In spring, solar radiation provides a strong downward flux of heat, the ice melts and any impurities contained in the ice are released into the water column. In small lakes, the covering of ice stops any transfer of momentum from the wind to the water body. In large lakes, the ice may break and give rise to some episodic movement. In shallow lakes the bottom sediment forms a significant store of heat that can influence their winter thermodynamics (e.g., Golosov et al., 2006). The volumetric changes associated with the formation of ice are of no consequence in deep lakes but in shallow lakes there may be substantial relative reductions in the volume of liquid water.

The ice studies organized in CLIME included observations of the heat fluxes across the ice sheet, the structure of the ice, the impurities in the ice sheet and the mathematical modelling of the growth and decay of the ice. The most intensive studies were conducted on Lake Pääjärvi, a relatively large lake in southern Finland. Here, an automatic monitoring station (Fig. 5.1) was installed to collect data on the structure of the ice, the thermodynamics of the lake and to calibrate the ice model (Wang et al., 2005). In Pääjärvi the ice season lasts 4–6 months (on average from November 30th to May 5th), and the ice thickness reaches its annual maximum of 30–80 cm in March (Kuusisto, 1994; Kärkäs, 2000; Leppäranta et al., 2006). Another lake, Lake Peipsi in Estonia and Russia, was used to test a two-dimensional mechanical model of the ice. Lake Peipsi has a surface area of 3,552 km^2 and its

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Fig. 5.1 The automatic station used to acquire data on the development of ice in Pääjärvi, the lake site used to test the new ice model

large size means that the ice cover is broken by winds and frequently drifts to form leads and pressure ridges.

Mathematical modelling of lake ice is approached by using a thermodynamic model for ice growth and decay and a mechanical model for ice displacements. Thermodynamical lake ice models have mostly been semi-analytical based on the freezing-degree-days (e.g., Ashton, 1986). A few investigators have developed numerical models, but even then the treatment of the slush and snow-ice formation has not been done in a rigorous way. Mechanical models have hitherto only been produced for the North American Great Lakes and then only some 25 years ago (e.g., Wake and Rumer, 1983). In CLIME, more recent experiences of modelling sea ice in small basins (Saloranta, 2000; Leppäranta, 2005) were used to revise both types of models of lake ice. The thermodynamic component of the new lake ice model consists of four realistically interactive layers – snow, slush, snow-ice and congelation ice – and the mechanical component employs a plastic ice rheology. In this model the ice is forced by the wind and its mobility is determined by the yield strength of the ice. If the forcing is large enough, the ice cover is shifted until a new equilibrium configuration is achieved.

This chapter gives an overview of the physics of lake ice and summarizes the results of the CLIME modelling studies. A complementary chapter, Chapter 4 in this volume, discusses the phenology of lake ice in relation to long-term changes in the climate. The present chapter is divided into four sections. In the first section the structure of lake ice cover is described in relation to its stability, stratigraphy, impurities, and optical properties. The second section discusses the growth and decay of lake ice and includes some heat flux results from an instrumented lake. The third section deals with the numerical modelling of ice growth and decay under both current and future climatic conditions. The fourth section presents some example results from the model simulations for Pääjärvi and Peipsi.

5.2 The Structure of Lake Ice

5.2.1 Stability

The classification of ice-covered lakes is here based on their mechanical properties. In small lakes, the ice cover is static, apart from fractures and thermal cracks. In large lakes the ice is mobile and may experience large wind-driven displacements (Fig. 5.2). In between, there is a grey area of ‘medium size’ lakes, where the ice cover is static in cold winters and may be mobile in mild winters. This distinction primarily depends on the size of the basin relative to thickness of the ice. Examples of large lakes with mobile ice cover are Lakes Peipsi, Ladoga and Vänern.

The ice cover on a lake can be viewed as a plastic medium forced by the wind stress $\boldsymbol{\tau}_a$ (Leppäranta, 2005). The critical point for a mechanical breakage event is whether

$$\tau_a/P^* > h/\ell \tag{5.1}$$

where $P^* \sim 30$ kPa is the strength of ice of unit thickness, h is ice thickness, and ℓ is the fetch, at most ℓ equals the basin size L. Wind stress for wind speed $U_a = 10$ m/s is $\tau_a \approx 0.3$ Pa (e.g., Andreas, 1998), and then $\tau_a/P^* \sim 1{\cdot}10^{-5}$.

In Lake Peipsi, $L \approx 50$ km and $h \sim 30$ cm, thus $h/L \sim 0.6{\cdot}10^{-5}$ and the ice is mobile. For a typical lake in southern Finland, $L \sim 10$ km and $h \sim 40$ cm, and the ice cover is stationary; but with major decrease in ice thickness and/or increase in wind speed the ice cover can become mobile. Very rare observations of modest ridging exist from the past in Lake Pääjärvi ($L = 10$ km) but in general the ice cover is static.

If the inequality (5.1) is satisfied, the ice motion is obtained from the quasi-steady momentum balance and ice conservation law

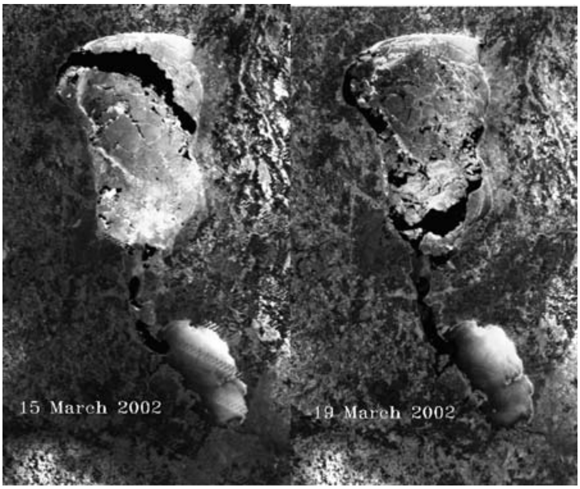

Fig. 5.2 Lake Peipsi ice cover on March 15th and 19th 2002 as shown by MODIS images acquired by NASA’s Terra/Aqua satellite. In the north a wide lead is seen in the 15th, and in four days the lead closed up accompanied by opening and fracturing in the south. The length of the lake is 150 km

$$\boldsymbol{\tau}_{\mathrm{a}} + \boldsymbol{\tau}_{\mathrm{w}} + \nabla \cdot \boldsymbol{\sigma} = 0 \tag{5.2a}$$

$$\frac{\partial h}{\partial t} + \boldsymbol{u} \cdot \nabla h = -h \nabla \cdot \boldsymbol{u} + \phi \tag{5.2b}$$

where $\boldsymbol{\tau}_{\mathrm{w}}$ is the shear stress at the ice–water interface, $\boldsymbol{\sigma}$ is the internal stress, t is time, $\boldsymbol{u}$ is ice velocity, and ϕ stands for thermodynamics changes. The motion of ice results in the opening of wide leads in divergent zones on the lee side and the formation of pressure ridges in compression zones on the windward side. As the ice moves, the ratio *h/L* increases, and the motion ceases as soon as inequality (5.1) is no longer valid. The inertial time scale of floating ice is less than one hour, and mechanical lake ice events are relatively rapid shifts rather than a long-term drifting process.

5.2.2 Structure

There are three principal layers in a static lake ice sheet (Fig. 5.3): primary ice, congelation ice and snow-ice (Michel and Ramseier, 1971; Gow and Govoni, 1983; Leppäranta and Kosloff, 2000). When a lake starts to freeze, ice crystals form at the surface and join to form a thin sheet of ice, the so-called primary ice. Freezing in calm conditions produces a very thin (millimetres thick) sheet of primary ice

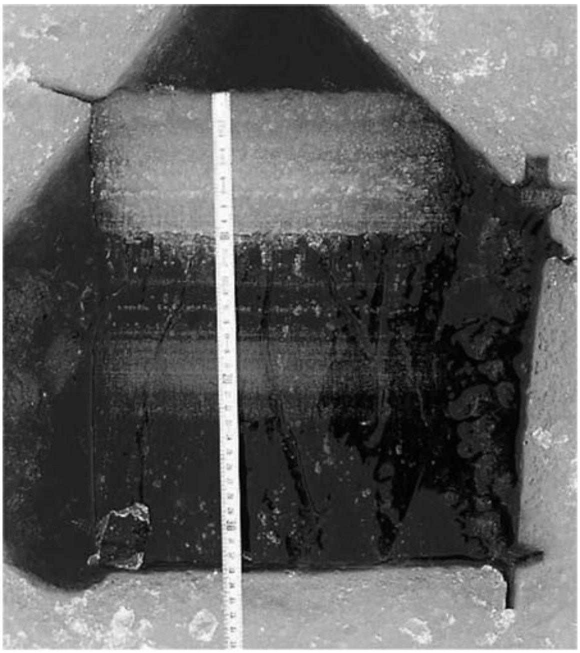

Fig. 5.3 A cross-section of an ice sample from Lake Pääjärvi (14 April 2004). Total thickness 33 cm (snow-ice thickness 10 cm and congelation ice thickness 23 cm)

where the optical axes of the ice crystals are aligned vertically. When the ice forms under more disturbed conditions, the primary ice layer is formed of frazil ice where the crystals are more randomly oriented. Congelation ice is formed as the ice sheet extends into the water column. If the crystals in the primary ice have a vertical orientation, the same pattern develops in the congelation ice with the formation of large macrocrystals. Otherwise, the congelation ice has a columnar crystal structure with optical axes turning horizontal. In both cases the crystal size increases with depth.

The ice on most Boreal lakes is covered with snow. Snow differs from ice in many respects. It is a good insulator and thus accentuates the decoupling of the water body from the atmosphere. The light transmitting property of snow is low compared with ice. Part of the snow cover may also change into slush by mixing with liquid water from precipitation, melting snow or local flooding. Snow-ice forms when this slush freezes. Since this progresses from the top down, pockets of slush can remain within the ice sheet and may persist for several weeks.

In Pääjärvi, the structure of the winter ice sheet has been monitored for 12 consecutive years (Table 5.1). The ice cover is usually static but in one winter when the ice was relatively thin (33 cm), a minor mechanical shift was observed with formation of a small ridge. The annual maximum ice thickness has ranged from 33 to 80 cm and the relative proportion of snow-ice has ranged from a minimum of 10% to a maximum of 40%. Usually the ice forms on cold calm nights in late autumn and produces a vertical orientation of the crystal axes and large macro-crystals in the congelation ice. Snow-ice forms mainly by flooding. In three winters out of the twelve a quasi-permanent slush layer several centimetres thick developed within the snow-ice layer. The total thickness of ice as well as the thicknesses of congelation ice and snow-ice layers is primarily controlled by the air temperature and snowfall

Table 5.1 Ice layers in Lake Pääjärvi (March). The thicknesses of congelation ice, snow-ice and snow, and water level elevation with respect to the ice surface (cm)

Year	Congelation ice	Snow-ice	Snow	Water level	Comments
1993	30	3	10	−0.1	Low ridge in the lake
1994	54	13	5	3.0	
1995	26	17	7	2.0	
1994	54	13	5	3.0	
1995	26	17	7	2.0	
1996	38	5	30	−5.0	
1997	21	14	1	3.0	Slush layer with algae in ice
1998	42	10	5	4.0	
1999	30	23	6	3.0	Slush 14–17 cm from top
2000	22	14	7	2.0	
2001	20	17	1	3.0	
2002	30	11	2	2.0	
2003	70	10	0	–	Freeboard estimated as 10 cm
2004	33	12	22	2.0	
2005	31	20	8	2.0	Slush layer with algae in ice

time history. A general trend toward warmer or colder winters may thus not be easy to detect in the ice thickness data.

In large lakes, frazil ice may form in leads and be incorporated into the adjoining ice sheet as frazil layers. The resulting stratification may show frazil layers and congelation layers in turn. This kind of ice structure is typical in freezing rivers and in seas where leads are common throughout the winter.

5.2.3 Impurities

Lake ice also contains impurities. Gas bubbles, several millimetres in size are common (Fig. 5.3). In congelation ice the gas is derived from the water column or the bottom sediments, while snow-ice contains enclosed air pockets from the parent snow. The particles trapped in the ice sheet may originate from the lake water, the bottom sediments or atmospheric fallout. Congelation ice is typically much cleaner than the water from which it forms. Frazil ice crystals capture particles from the liquid water whilst snow-ice can include impurities from the lake water in the parent slush. When the ice melts, these impurities are released into the lake in a very short time. The meltwater of clean congelation ice may then be less dense than the underlying lake water due to its very low content of dissolved matter.

In Lake Pääjärvi, a number of meltwater samples have been examined for the presence of impurities (Leppäranta et al., 2003b). The mean conductivity of the meltwater was 12 μS/cm (at 25°C), which represents 10% of the conductivity of the lake water. The average dissolved matter content was 14 mg/l, whilst the concentration of suspended matter was 1.9 mg/l, which is also less than that recorded in the liquid water. Most of the ice impurities appeared to be the result of the flooding of lake water on the ice and atmospheric fallout. When congelation ice forms in fresh water, the ice-water interface is planar and dissolved matter is to a large degree rejected from the ice (Weeks, 1998). Congelation ice does not provide an appropriate environment for living organisms, but algae can grow in a slush layer within snow-ice. If the total ice thickness were to decrease under climate warming, the buoyancy of ice would be less and flooding events would become more common.

5.2.4 Optical Properties

Ice cover has a major effect on the transfer of sunlight into the water column (Leppäranta et al., 2003a; Arst et al., 2006). The transmission of light can be modelled using the Beer's law $dq/dz = -\mu q$, where $q = q\,(z;\,\lambda)$ is irradiance, z is depth, λ is wavelength, and $\mu = \mu(z;\,\lambda)$ is the attenuation coefficient. The general solution is

$$q\,(z;\lambda) = (1 - r)q_o\, exp\left(-\int_0^z \mu dz'\right) \tag{5.3}$$

where $r = r\,(\lambda)$ is surface reflectance and $q_o = q\,(0;\,\lambda)$. The reflectance ranges from 0.07 for open water to 0.2–0.5 for ice and 0.5–0.9 for snow. Therefore the level of net incoming irradiance into a lake is drastically reduced by ice cover. Only the optical and ultraviolet components of the spectrum have any significant penetration, and their total level follows the attenuation law in Eq. (5.3).

Light attenuation in clear congelation ice is very close to that in liquid lake water, i.e. the attenuation spectra are similar and the attenuation coefficient $\mu \sim 1\ m^{-1}$ (Fig. 5.4). Since the concentration of impurities is lower in congelation ice than in the lake water, the ice may be even more transparent than the water in turbid or humic lakes. In contrast, gas bubbles in the ice scatter light which lowers the transparency and flattens the attenuation spectrum. Snow-ice and snow appear opaque and white due to the large volume of the air inclusions and have the optical depth close to 15–20 cm (Wang et al., 2005). For a layered ice sheet, the transparency is given by:

$$\tau = q_b/q_o = (1 - r)\exp[-(\mu_s h_s + \mu_{si} h_{si} + \mu_i h_i)] \tag{5.4}$$

where q_b is the downwelling irradiance just beneath the ice, and h is the thickness of sublayers with the subscripts i, s and si for ice, snow and snow-ice, respectively. For bare congelation ice the penetration is lowered due to the increased surface reflectance but a covering of snow makes a big difference due to its very high reflectance and rapid light attenuation: e.g., adding 20 cm snow on a 50 cm ice sheet can reduce the irradiance level beneath the ice by an order of magnitude.

In spring, sunlight plays a major role in the melting of the ice. It heats the ice and water, causes internal melting, and triggers the onset of the spring growth of phytoplankton beneath the ice. As soon as the snow has melted, internal melting of ice and warming of the underlying water begins. At ice break-up, the bulk temperature of the water below the ice can be as high as 4°C leaving just a thin layer of cold water beneath the ice. Under these conditions, the spring overturn may be very

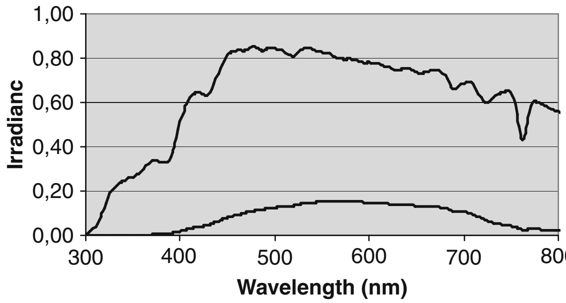

Fig. 5.4 Spectral attenuation of light by ice cover in Pääjärvi, 14 April 2004. The spectral irradiance is given in $W\ m^{-2}\ nm^{-1}$. Upper curve–downwelling irradiance at the surface; lower curve–downwelling irradiance just beneath the ice

short or even absent in some extreme years. Internal melting is normally strong, and in consequence springtime ice becomes porous.

The euphotic depth is usually taken as the depth where the level of irradiance is 1% of that recorded just below the surface (see Smetacek and Passow, 1990; Arst, 2003). In open water conditions, this is given by $(1-\alpha)\log(100)/M_w$, where α is albedo and M_w is the total light attenuation coefficient in water. In the presence of ice, the irradiance at the water surface is reduced and the euphotic depth becomes $(1-\alpha)\log(100T)/M_w$, where T is the total light transparency of the ice. Because of the albedo effect, the euphotic depth beneath bare ice is about half of that in the open water. If there is more than 20 cm snow on the ice no photosynthesis is possible in the upper layers of the water column.

5.3 The Growth and Decay of Lake Ice

5.3.1 Ice Thermodynamics

In the growth of congelation ice, latent heat of freezing is released at the bottom of the ice sheet and conducted through the ice into the atmosphere. The growth of the ice sheet continues as long as the conductive heat flux through the ice is greater than the heat flux from water to ice. The thicker the ice, the greater the distance for this conduction and the slower the subsequent rate of ice growth. In the presence of slush, snow-ice forms in the upper layers in a similar way and since slush contains ice crystals less latent heat is released. The melting season begins, when there is a positive balance in the net radiation. Once the temperature of the ice-sheet is uniform (0°C) melting takes place at the boundaries and by the absorption of solar radiation inside the ice sheet. Internal melting gives rise to structural defects and once the porosity of the ice reaches 0.3–0.4 the ice cannot bear its own weight and there is a rapid increase in the rate of decay.

The growth and melting of lake ice are vertical processes. Horizontal variations can occur but on length scales of more than a few meters the thermodynamic processes are essentially independent. In this situation, the equation for the conduction of heat becomes:

$$\frac{\partial \rho c T}{\partial t} = \frac{\partial}{\partial z}\left(\kappa \frac{\partial T}{\partial z} - Q_s\right) \tag{5.5a}$$

$$Surface{:}\kappa \frac{\partial T}{\partial z} = Q_o + m(T)\rho L \frac{dh}{dt} \tag{5.5b}$$

$$Bottom{:}T = 0^{\circ}C, \kappa \frac{\partial T}{\partial z} = Q_w + \rho L \frac{dh}{dt} \tag{5.5c}$$

where c is specific heat of ice, T is temperature, κ is thermal conductivity, Q_o is the net heat flux at the upper surface, Q_w is the heat flux from the water, and $m(T) = 1$ for $T = 0$°C or zero otherwise. Equations (5.5b, c) state that the heat fluxes are continuous through the upper and lower surfaces: conduction into ice equals external heat flux plus heat release or take-up due to phase changes.

5.3.2 Boundary Conditions

The variation in the incoming solar radiation and the exchange of mass and heat with the atmosphere define the boundary conditions at the surface of the ice-sheet. They are given, respectively, as

$$h' = P - E \tag{5.6a}$$

$$Q_0 = (1-\gamma)[Q_s(0) - Q_r] + Q_{L\downarrow} - Q_{L\uparrow} + Q_C + Q_e + Q_P \tag{5.6b}$$

where P is precipitation and E is evaporation, γ is the fraction of solar radiation (Q_s) penetrating the surface, Q_r is outgoing solar radiation, $Q_{L\downarrow}$ and $Q_{L\uparrow}$ are incoming and outgoing terrestrial radiation, Q_c and Q_e are sensible and latent heat fluxes, and Q_P is the heat provided by precipitation. The radiation terms are given by:

$$Q_s(0) = \Gamma(N, e, \theta)Q_{so}, Q_r = (1-\alpha)Q_s \tag{5.7a}$$

$$Q_{L\downarrow} = (a + be^{1/2})(1 + cN^2)\varepsilon\sigma \underline{T}_a^4, Q_{L\uparrow} = -\varepsilon\sigma \underline{T}_o^4 \tag{5.7b}$$

where Γ is atmospheric transmissivity, N is cloudiness, e is water vapour pressure, θ is solar zenith angle, Q_{so} is the solar constant, ε is emissivity of the ice/snow surface, σ is Stefan-Boltzmann constant, T_a is air temperature, $a= 0.68$, $b = 0.0036$ mbar$^{-1/2}$ and $c = 0.18$ are empirical parameters and the underlined temperatures refer to the absolute temperatures. The bulk formulae have been used for the turbulent fluxes:

$$Q_c = \rho_a c_p C_H (T_a - T_o) U_a, Q_e = \rho_a L_E C_E (q_a - q_o) U_a \tag{5.7c}$$

where ρ_a is air density, c_p is specific heat of air at constant pressure, L_E is latent heat of evaporation, q_a and q_o are specific humidity of air and surface, and C_H and C_E are the bulk exchange coefficients (C_H, $C_E \sim 1.5 \cdot 10^{-3}$). The heat flux from the precipitation is:

$$Q_P = [\rho c (T_P - T_o) + \Pi \rho L] \mathrm{d}P/\mathrm{d}t \tag{5.7d}$$

where T_P is the temperature of the precipitation and $\Pi = -1$, 0 or 1 if the phase of the precipitation changes from solid to liquid, there is no phase change, or the phase of the precipitation changes from liquid to solid. This heat flux may be important if there are phase changes, otherwise it is negligible.

The surface heat budget (Fig. 5.5) was derived from the high-resolution measurements recorded by the Automatic Ice Station. The radiation balance (solar plus terrestrial radiation) dominates the total heat flux. The level of the turbulent loss

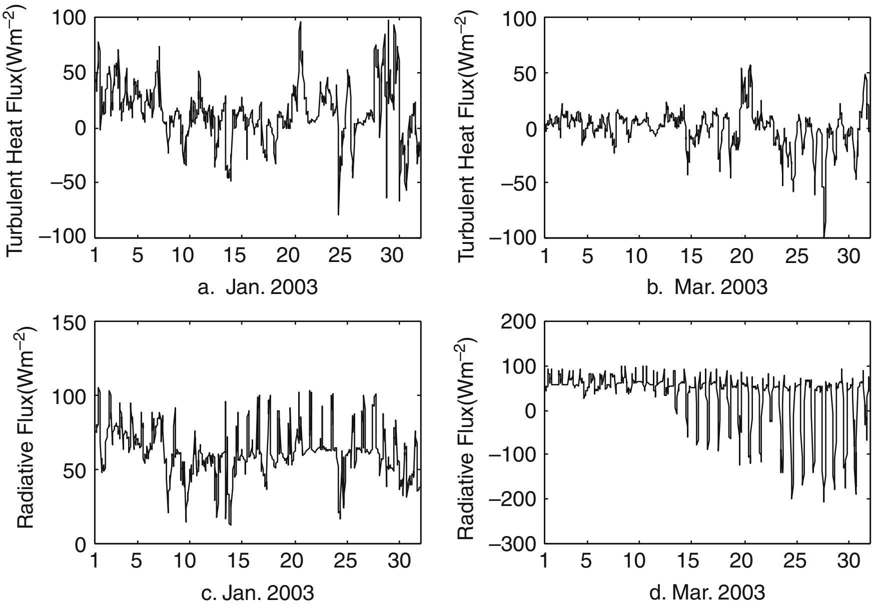

Fig. 5.5 Heat fluxes calculated from the data acquired by the Pääjärvi ice station during the winter of 2003 (Wang et al., 2005): a – turbulent heat flux in January, b – turbulent heat flux in March, c – radiative heat loss in January, and d – radiative heat loss in March

was only high for short periods and typically ranged between 20 and 30 W/m^2 in January (Fig. 5.5a) and between −10 and 10 W/m^2 in March (Fig. 5.5b). In January it was 50–70 W/m^2 due to the strong outgoing terrestrial radiation from the surface (Fig. 5.5c), and at about mid-March the daytime solar radiation was high enough to turn the balance into a gain of up to 200 W/m^2 (Fig. 5.5d).

At the bottom of the ice sheet, the heat balance is determined by phase changes, the conductive flux of heat into the ice sheet and the transfer of heat from the water column. In the case of laminar flow, the water–ice heat flux is $Q_w = k_w\, dT/dz$, $k_w = 0.6$ W/(m °C) being the molecular conductivity of heat for liquid water. For $dT/dz \sim 1$°C/m, we thus have $Q_w \sim \frac{1}{2}$ W/m^2. In the case of turbulent flow, the bulk formula is

$$Q_w = \rho_w c_w C_{Hw} T_w U_W \tag{5.8}$$

where ρ_w is water density, c_w is the specific heat of water, C_{Hw} is the heat exchange coefficient, T_w is water temperature, and U_w is current speed. Lake Pääjärvi station data gave the estimate of $C_{Hw} = 0.4 \cdot 10^{-3}$ and the heat flux magnitude of 2–5 W/m^2 (Shirasawa et al., 2006).

5.3.3 Numerical Modelling of Ice Growth and Decay

The traditional semi-analytical modelling approach to the growth and decay of lake ice has severe limitations. First of all, the snow cover cannot be properly considered. The insulating capacity of snow is primarily dependent on snow accumulation, melting and metamorphosis which can even lead to slush and snow-ice formation. The resulting lake ice has then a multi-layer structure. Also a realistic air−ice interaction modelling necessitates high-resolution information of the ice temperature structure. Numerical models of lake ice thermodynamics are based on the solution of the heat conduction law together with phase changes (Eqs. 5.5, 5.6, 5.7, and 5.8) using a grid size 1–5 cm. However, the ice growth modelling problem possesses a negative feedback to errors. The background Stefan's law implies that squared ice thickness is proportional to the freezing-degree-days S, and consequently the sensitivity of ice thickness reads $\delta h \propto \delta S/h$ (Leppäranta and Kosloff, 2000). Therefore, the total thickness of the ice is quite well represented by semi-analytical models, but the internal structure of ice is not resolved and sensitivity of thickness to various factors is unclear.

Numerical congelation ice models with passive snow layer were first developed in the 1970s (Maykut and Untersteiner, 1971) and later extended to include an interactive snow model (Leppäranta, 1983; Saloranta, 2000; Shirasawa et al., 2005). In the CLIME project, a new lake ice model has been developed based on the earlier sea ice models and our recent lake research results (Leppäranta and Uusikivi, 2002). The new feature is that the phase proportions are simulated for each grid cell in addition to the temperature. This approach allows a realistic structural profile with locations of liquid water containing layers. The model is forced by heat fluxes from air and water and the mass flux from precipitation. The details of the CLIME ice model will be published elsewhere (Leppäranta, 2009).

The full model simulates the development of four distinct layers: snow, slush, snow-ice and congelation ice. These layers, in turn, interact in a dynamic way: thus snow accumulation creates slush and snow-ice depending on the total thickness of ice, while the growth and decay of congelation ice depend on the snow and slush conditions. Slush and snow-ice may form a multiple layer structure with alternating layers of snow-ice and slush. The snow layer needs its own model. The thickness of snow decreases due to three different reasons: surface melting, compaction, and formation of slush, which further transforms the accumulated snow into snow-ice. The change in the density of snow depends mainly on the mass of overlaying snow and temperature and the thermal conductivity is proportional to the density squared (Yen, 1981). The model was forced by Lake Pääjärvi Ice Station data for the calibration. These data were collected by the Pääjärvi Automatic Ice Station with a time resolution of 10 min (see the previous section).

Figure 5.6 shows the outputs generated by the model for the ice season 2002/2003 starting from the day when the station was deployed (December 17th 2002). In the beginning the ice was bare with a thickness of 19 cm. The model produced a good fit to the observations until the ice reached its maximum thickness of 80 cm in March i.e. the difference between the model and the observations was always less than

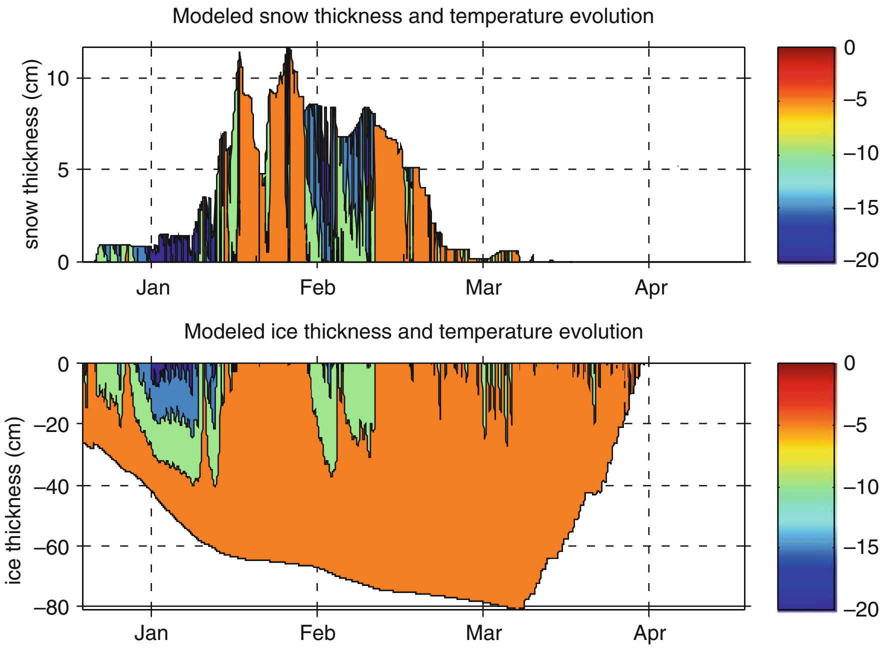

Fig. 5.6 Thermodynamical model simulation of the ice and snow thickness for January − April 2003 in Lake Pääjärvi (Leppäranta et al., 2006). For clarity the near-zero temperatures are shown by one colour

5 cm. There were two very cold periods, in the first half of January and the first half of February, when the ice grew fast, at a maximum rate of 2 cm per day. In the first period the snow cover was still thin, and the temperature in the surface layers of the ice was below −20°C. In the model, the thickness of ice increased to 65 cm by the end of January. In February there was some snow accumulation and ablation, but the snow thickness was generally less than 10 cm. The melting season in the model appears rather too early (on March 10th) and all the ice had melted by the beginning of April, two weeks earlier than observed. This result is not particularly good, but can be understood on the basis of the problem of the parameterisation of the most critical factor − the albedo. Here the snow layer is modelled on the basis of precipitation input as liquid water equivalent, and so small errors can have a significant effect on the onset of melting in the model. In the melting stage, however, the rate of melting in the model agrees well with the observations.

The annual cycle of the change in ice thickness is well represented by the model and demonstrates that the model can be used to simulate the evolution of the ice sheet under a range of meteorological conditions. The weakest components of the model appear to be snow formulation together with the onset of melting, which need further testing with winter seasons of varying snow conditions. The heat flux from the water to the ice is fixed in the model; it has been estimated as 2–5 W/m^2 in Pääjärvi (Shirasawa et al., 2006), which decreases the annual maximum thickness by about 5 cm. The heat flux from the lake bottom to the water column also needs to be considered especially in shallow lakes.

5.4 Climate Change and Ice Conditions

5.4.1 The Characteristics of Ice Seasons

An ice season is characterized by the freezing and ice break-up dates, the length, and the maximum annual ice thickness. Time series analyses (see Chapter 4 in this volume) and analytical models (below) can be used to provide the first-order predictions for the sensitivity of these fundamental characteristics of the ice season to climate variations. In the first order modelling approach the surface heat balance is taken in general form $R_o(t) + r_1[T_a(t) - T_o(t)]$, where r_1 is a constant. Then a constant change ΔT_a can be employed for the air temperature level to examine the influence of an assumed climate change (Leppäranta, 1989; Haapala and Leppäranta, 1997). This first-order approach is quite appropriate for examining the general variations in the ice season since the simplifications in the surface heat budget are largely cancelled i.e. the freezing date estimate is rather crude but we get better estimates of the change of the freezing date for a given climate scenario. Here the change of the characteristics of ice seasons is examined in the light of an arbitrary (small) air temperature increase of ΔT_a, numerical examples with $\Delta T_a = 1°C$. The prediction for a given scenario is then easily evaluated from the derived general formulae.

The date of freezing is dictated by the heat loss from the lake to the atmosphere and the heat that the lake accumulated in summer. Using a slab water body model and a fixed atmospheric cooling rate $\dot{T}_a$ in the autumn, the freezing-day becomes delayed by

$$\Delta t = \Delta T_a / \dot{T}_a \tag{5.9}$$

(Leppäranta, 1989). In southern Finland, $\dot{T}_a \approx 5°C/\text{month}$, and consequently for $\Delta T_a = 1^{\circ}C$ we have $\Delta t = 6$ days.

At the edge of the geographical area where lakes freeze in winter the projected changes are more qualitative i.e. a lake that now freezes in winter may become ice-free. Thus, the critical question is whether a lake freezes or not. Using a parabolic curve for the change in the winter air temperature, we can derive the following equation from the slab model: a lake freezes if $t_f > 2t_r$, where t_f is the length of the period the air temperature is below the freezing point and t_r is the response time of the lake. For an air temperature change ΔT_a the criterion changes into

$$t_f\sqrt{1 + \Delta T_a / \min(T_a)} > 2t_r, \tag{5.10}$$

$\min(T_a) < 0$. The response time depends first of all on the depth of the lake, and $t_r \sim H\ \text{m}^{-1}$ d; e.g., for $H = 10$ m we have $t_r \sim 10$ d.

So-called freezing-degree-day models are used for the analytic ice growth laws (see Leppäranta, 1993). Zubov's (1945) law is $h = \sqrt{aS + d^2} - d$, where $a \approx 11$ cm^2 °C^{-1} day^{-1}, S is the number of freezing-degree-days, and d corresponds to the insulating effect of the near-surface air-snow buffer. Lammi station data give normally annual freezing-degree-days of 300–400°C·day. Zubov's law would give estimates of 48–57 cm for the maximum annual ice thickness, about 5–10 cm more than in reality. The result for a selected rate of climate change is

$$\Delta h = \sqrt{aS_2 + d^2} - \sqrt{aS + d^2} \tag{5.11}$$

The relationship is non-linear, and therefore rate at which the thickness changes depends on the initial level. For a parabolic form of the evolution of the winter air temperature, we have $S = -2/3\min(T_\mathrm{a})t_\mathrm{f}$. If $\min(T_\mathrm{a}) = -10$°C and $t_\mathrm{f} = 100$ days, the decrease in the thickness of the ice is from 76 to 70 cm for $\Delta T_\mathrm{a} = 1$°C; for $\min(T_\mathrm{a}) = -5$°C and $t_\mathrm{f} = 50$ days, the corresponding drop would be from 43 to 35 cm. As a first order approximation, the growth of ice is proportional to the square root of the freezing-degree-days. Thus the sensitivity of ice thickness to temperature conditions is greater when the ice is thin. As t_f approaches $2t_\mathrm{r}$, most of the freezing-degree days is used into the cooling of the water, and the change in thickness becomes even greater.

The break-up date depends on three factors: the time of the onset of melting, the thickness of the ice and the weather conditions during the melting phase. The first factor is difficult to estimate but it is close to the zero upcrossing of the radiation balance. Thereafter, the melting of snow and ice is proportional to the positive degree-days, with the proportionality factor of 0.1–0.5°C·day/cm, depending on the albedo. The shift in the break-up date (Haapala and Leppäranta, 1997) thus becomes

$$\Delta t = \sqrt{\frac{2h_\mathrm{max}}{b(1-\alpha)}} - \sqrt{\frac{2h_\mathrm{max} - \Delta h}{b(1-\alpha)}} \tag{5.12}$$

where h_max is the maximum annual ice thickness for the current climate and $b = \dot{Q}_s/(\rho L)$ is the rate of increase in spring solar radiation scaled by the latent heat of melting. Taking normal ice thickness as 45 cm and melting period as about 45 days, this means that the break-up date will be 1 day earlier for every centimetre of ice lost, i.e. a loss of 5–10 cm will results in an advance of 5–10 days.

5.4.2 Numerical Modelling of the Seasonal Ice Cycle

In the final step, the numerical model was used to produce some projections for the future ice seasons. The baseline ice season is obtained with a simulation of a normal ice season in the current climate. Then a fixed shift is introduced into the air temperature data and the model is run for these new 'warm world' conditions.

According to the climate scenario produced by CLIME for Pääjärvi (Chapter 2 in this volume) the winter air temperatures projected for the area for 2070–2100 could be up to 6°C higher than they are today. This brings the mean winter temperature close to 0°C, which would have a dramatic effect on the winter characteristics of the lake.

Figure 5.7 compares the model simulations for a ‘normal’ winter with those for winters where the mean air temperatures are 1°C, 4°C and 6°C higher than this historical baseline. The results show that the 1°C change brings the thickness curve down by 5–10 cm but has no effect on the form of the thickness cycle. With a 4°C warming there are growth and decay periods through the winter and the maximum ice thickness is about 20 cm. In southern Finland the thickness of 20 cm is enough for a stable ice cover in small lakes, but in large or medium-size lakes stormy winds could break this cover and give rise to areas of open water at any time during the winter. When the temperature increase is 6°C very little ice remains on this lake. The length of the ice season is only 20–25 days, the maximum thickness is 11 cm and the ice cover is also mechanically unstable throughout the winter.

The numerical model developed in CLIME is a pure ice model based on a simple slab model for the lake water body. The surface layer is allowed to warm and cool to reproduce the effects associated with several freezing – melting events. The actual date of first freezing is not therefore predicted with any greater accuracy than the analytic model described above. The key feature of the scenarios presented here is that the rate of cooling in the autumn does not change but is just delayed. This means that, each 1°C increase in the air temperature delays the date of freezing by six days. Thus the projected 6°C warming would result in a delay of 36 days. The freezing criterion (Eq. 5.10) implies that the length of the period when the air temperature is below zero must be at least one month for freezing of Lake Pääjärvi.

In the numerical model, the lake freezes during the first strong cold spell. The observed maximum annual ice thickness of 45 cm decreases by 5 cm for the first 1°C warming, 20 cm for the next 3°C warming, and 10 cm for the final 2°C increase. The ice break-up date also shifts to a much earlier date due to the earlier onset of

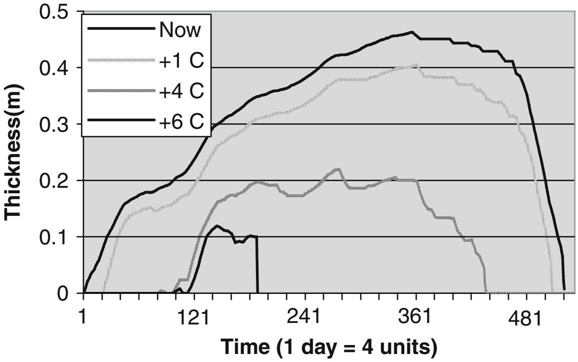

Fig. 5.7 Thermodynamical model simulations for the ice thickness for a normal winter in the present and warmer climates with temperature higher by 1°C, 4°C and 6°C. In time axis tic spacing is five days

melting which reflects the change in the radiation balance and the fact that there is less ice to melt. The resulting ice break-up is five days earlier for the 1°C warming, 25 days earlier for the 4°C warming, and 75 days earlier for the 6°C warming. This illustrates the high instability of the ice season for strong warming. In a warm climate, ice may form in mid-winter but since it is relatively thin it soon melts when the weather becomes warmer.

The projected changes in the climate will have a significant effect on the qualitative characteristics of the ice as well as the quantitative changes described above. In geographical terms, the ice season regimes can be grouped into three zones: ephemeral ice zone, unstable ice zone, and stable ice zone. In the first zone, ice comes and goes in one season, in the second the ice breaks at times and results in periods when the lake is partially open while in the third zone the whole lake has stable ice cover from the date of first freezing until the ice cover becomes rotten in spring. The boundaries of the zones will move north when the climate becomes warmer. The extent of the ephemeral zone is dictated by the mean and variance of air temperature whilst that of the unstable zone is dictated by the thickness of ice. With decreasing ice thickness, the period of the ice being breakable increases and finally the ice cover is breakable during the whole ice season. Today, only very large lakes at around 60°N in European climate have unstable ice cover but if the ice thickness is reduced to 20–30 cm or less, medium size lakes will also be unstable.

Consequently the implications of a warming climate to the lake ice season are first shorter season and then thinner ice. However, depending on the snowfall conditions, snow-ice formation may increase to counter at least part of the effect of climate warming. Thinner ice also means more breakable ice with more areas of open water which changes fluxes of heat and moisture between the lake and the atmosphere and may generate frazil ice. Frazil ice formation may, in turn, lead to accumulation of sediments in the ice cover via frazil capturing of particles in the water body and via anchor ice formation. From the point of view of lake ecology, shorter ice season and presence of openings mean fewer oxygen deficit problems. But the openings will also create a new type of winter environment in lakes, which have hitherto been completely covered with ice.

5.4.3 Mechanics of the Ice Cover

The application of the Lake Peipsi mechanics model has been described by Wang et al. (2006), see also Wang et al. (2003). It is based on the numerical solution of Eq. (5.2). The model was validated by comparing the physical changes simulated by the model with a series of MODIS images acquired by the Aqua/Terra satellite (Fig. 5.2). The first image shows that there was an open lead in the northern part of the lake on the 15th of March. The ice then drifted towards the northeast gradually closing the lead as more open water appeared in the southern and western parts. Another lead opened in the middle of the lake on the 18th March before the ice drifted towards the northwest. In the model simulation the initial ice thickness and

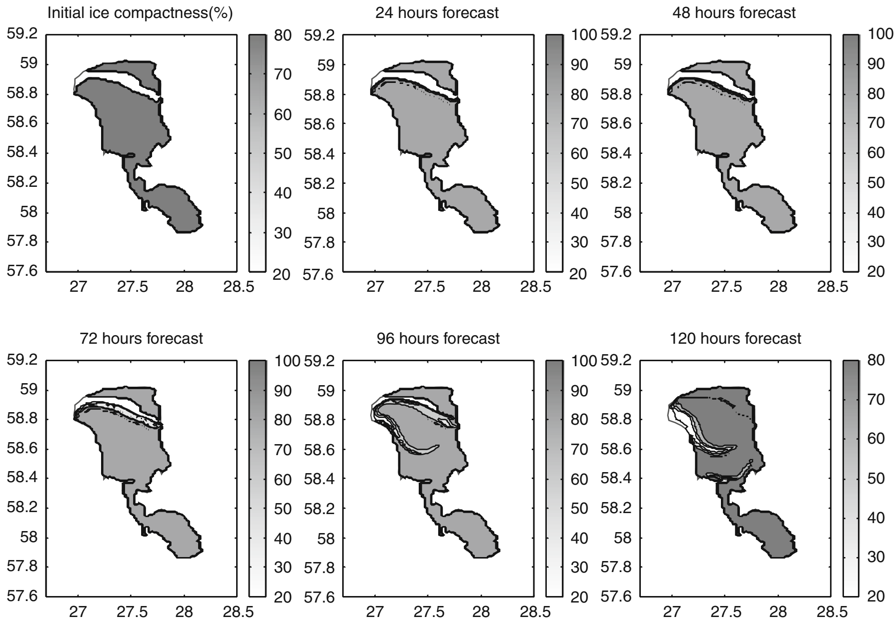

Fig. 5.8 Maps showing the results of the mechanical model simulation that showed the day-to-day variation in the compactness field for ice on Lake Peipsi between the 14th and 19th of March 2002 (Wang et al., 2006)

compactness were taken as 40 cm and 99%, respectively, for the ice areas. The winds were taken from the NCAR reanalysis data at 10 m altitude. Northern winds were dominant before 17th of March and southern winds thereafter. The highest wind speed was less than 7 m/s.

The model outputs (Fig. 5.8) follow the same pattern as the satellite images. In the first two days, the ice cover was immobile forced by the northern wind but a lead appeared in the middle part of the lake and on the 18th of March. The simulation for the following day shows the ice opening towards the south. The openings clearly result from wind forcing and the geometry of the lake shore. Thus, as the mechanical forcing reaches the yield strength of the ice cover, the ice seasons change from stable to unstable regime. Mechanical movements of this kind are very rare in Finnish lakes. The mobility of the ice cover is given by the criterion $h/\ell < \tau_a/P^*$ (see Section 5.2). In southern Finland, the thickness of lake ice is currently 30–60 cm, and even with long fetches $\ell \sim 10$ km the minimum thickness still results in stable ice cover. With this fetch and a wind speed of 15 m/s, the ice would have to be ca 15 cm thick for any significant breakage to occur. Since 1°C climate warming would reduce the ice thickness by 5–10 cm, ice breakage would become common in Finnish lakes as soon as the winter air temperature has increased by 2–3°C.

5.5 Final Remarks

Ice cover plays a major role in the seasonal dynamics of boreal and mountain lakes. The ice and snow restrict the exchange of momentum and heat between the atmosphere and the water, limit the vertical transmission of light and have a major effect on the way the lakes are used by the local population. This overview of the physics of lake ice is based on a review of the literature and an analysis of the new high-resolution data acquired by CLIME. This includes the structure, optical properties and stability of lake ice cover and the physical processes that influence the growth and melting of lake ice. The chapter includes a brief description of the new mathematical lake ice model developed in CLIME and considers the potential impact of climate change on the ice conditions in European lakes. The results presented demonstrate that the projected changes will have a profound effect, not only on the length of the ice season and the thickness of the ice, but also on the mechanics of the ice and lake ecology.

The lake ice modelling problem can be approached with semi-analytic or numerical methods. Semi-analytic methods are based on the degree-days and provide the first order approximation of the freezing – ice growth – melting cycle from which simple relations between winter climate and ice season characteristics can be derived. Numerical models are designed to include the full ice physics in the analysis. Numerical models of this kind have been widely used in thermodynamic studies of sea ice but have not hitherto been used for freshwater investigations. In these models the stratigraphy of ice, snow physics and atmospheric forcing can be treated in a realistic way, and the evaluation of climate change consequences becomes more reliable than with semi-analytic models. The CLIME ice thermodynamic model is based on the existing sea ice models, and the new feature is the inclusion of the liquid water content of the ice and snow as an additional variable. This means that slush layers are well reproduced and the physical representation of melting ice is more realistic. In addition, we also developed a mechanical model that could be used to examine the breakage of ice, a factor, which is a critically important to the character of lake ice seasons. Mechanical models have not been much discussed in the lake ice literature over the last 25 years. Lake ice mechanical events are thickness dependent and therefore particularly sensitive to climatic variations.

The CLIME ice model proved an effective tool to evaluate the influence of climate change on the ice conditions of frozen lakes. The resulting ice cycle depends on the atmospheric mass and heat fluxes and is particularly sensitive to air temperature and snowfall. The connection with the ice and snowfall is quite complicated since the timing of the snowfall is critical. Also the ice break-up date depends on the structure and thickness of the ice and snow and needs a physically realistic model, with stratigraphy, to explore its variability. The sensitivity of ice thickness in models is in general inversely proportional to its thickness. Therefore when the ice is thin we need a more sophisticated model to simulate the freezing and thawing.

In CLIME, the climate change scenarios used for the regional assessments were based on the outputs of two Regional Climate Models driven by two established emission scenarios (see Chapter 2 this volume). For Lake Pääjärvi, the key climatic variable was the change in the winter air temperature, which was projected to increase by as much as 6°C by 2070–2100. This degree of warming naturally results in a much shorter ice season and thinner ice. In the present climate, the average length of ice season in southern Finland is 4–5 months and the maximum annual ice thickness is, on average, close to 40 cm. The results presented here demonstrate that a warming of as little as 2–3°C would result in major qualitative changes. In most lakes, the ice is would then be thin to be unstable, be breakable by winds to produce extended periods with open water at any time during a winter. The full 6°C warming scenario would result in an average ice season of 20–25 days and maximum ice thickness of about 10 cm. If the warming trend is progressive, this means that the characteristics of the ice season in these lakes would be dramatically different by 2050.

In these cold regions people have learned to live with frozen lakes and utilize them to their advantage. In the past, lake ice was used to store perishable products, when it was removed during the winter and stored for the summer. Special fishing techniques have also been developed for ice-covered lakes, and the ice cover serves as a good base for winter roads. Frozen lakes are also used as a venue for many recreational purposes, such as ice fishing, skiing, long distance skating and ice sailing. On the negative side, the freezing of lakes makes boating impossible and limits their use for commercial and recreational transport.

It is clear that shorter ice seasons would severely limit these traditional activities. On the other hand, the lakes would be open for a much longer period and allow the extension of many water-based activities. For the ice cover to be 'useful', it has to be thick enough to serve as a solid platform. Lakes in the ephemeral zone, and to a lesser extent those in the unstable zone, are not useful in that sense. The bearing capacity (P) of floating ice is known to be $P = ch^2$, where $c \approx 5$ kg/cm^2. For a skier or skater 5 cm ice thickness is enough, but for using cars or snowmobiles the thickness should be at least 20 cm. When the thickness of the ice is very close to these minima special care must be taken because of natural variability of ice thickness and the resonance effects with shallow water waves for moving loads. If the ice is thicker, it will also be stable in medium size lakes: 30 cm is commonly taken as the safe thickness for recreational purposes and for traffic with small vehicles that weigh less than 1 ton. For trucks or trains the limiting thickness is very much greater.

A key point is thus the length of transition period from open water to 'useful ice', i.e. the period when the thickness of the ice is below the critical level of about 30 cm. Even a modest warming will have a major impact on the ice season viewed from this practical perspective. A warming by 2–3°C could reduce the ice thickness below the critical level throughout southern Finland and make the remaining ice cover dangerous throughout the winter. If there were a progressive increase in the projected rate of change, these qualitative changes would become very obvious in the closing decades of the century.

The physical factors influencing ice cover and its sensitivity to climate change are now quite well understood. The impact of these changes on the ecological status of lakes is less clear and must be the topic of the next step of investigations. This will require the analysis of ice cover in two-dimensions with appropriate process-based links to the seasonal dynamics of the lakes. Key questions include: the physical implications of the fracturing of lake ice and the effects that the projected change in the quality of the ice will have on the transfer of light. Even more important will be the construction of coupled ice and liquid water body models for boreal lakes. The coupling of the ice and water takes place via the solar radiation and heat loss, which both depend on the thickness of the ice. When a lake is frozen, the circulation in the water column is weak and driven by subtle variations in the flux of light and heat through the top and bottom surfaces of the lake. These weak circulations are, however, known to be ecologically important since they influence the oxygen budget, regulate the growth of phytoplankton and have a major effect on the winter survival of lake fish.

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Chapter 6
The Impact of the Changing Climate on the Thermal Characteristics of Lakes

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6.1 Introduction

Meteorological forcing at the air-water interface is the main determinant of the heat balance of most lakes (Edinger et al., 1968; Sweers, 1976). Year-to-year changes in the weather therefore have a major effect on the thermal characteristics of lakes. However, lakes that differ with respect to their morphometry respond differently to these changes (Gorham, 1964), with deeper lakes integrating the effects of meteorological forcing over longer periods of time. Other important factors that can influence the thermal characteristics of lakes include hydraulic residence time, optical properties and landscape setting (e.g. Salonen et al., 1984; Fee et al., 1996; Livingstone et al., 1999). These factors modify the thermal responses of the lake to meteorological forcing (cf. Magnuson et al., 2004; Blenckner, 2005) and regulate the patterns of spatial coherence (Chapter 17) observed in the different regions (Livingstone, 1993; George et al., 2000; Livingstone and Dokulil, 2001; Järvinen et al., 2002; Blenckner et al., 2004).

In this chapter, we summarise the long-term thermal changes observed in a number of lakes distributed throughout Northern, Western and Central Europe. These analyses complement the ice phenology results presented in Chapter 4, the ice modelling results in Chapter 5 and the temperature modelling results in Chapter 7. Particular attention is paid to the interannual and seasonal variations in the surface and bottom temperatures of the lakes. In Europe, lake surface waters are typically at their warmest in July or early August. Surface water temperatures in low-altitude lakes in central Europe then often exceed 25°C (Livingstone and Lotter, 1998; Livingstone and Padisák, 2007), but are usually lower in lakes at high altitudes or high latitudes (Livingstone et al., 1999; Korhonen, 2002; George et al., 2007b).

Long-term water temperature records are available from lakes in several different regions (e.g. Livingstone, 1993; Bengtsson et al., 1996; George et al., 2000; Livingstone and Dokulil, 2001; Nõges, 2004). These long-term data sets are

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especially valuable for evaluating the thermal responses of lakes of varying size, topography and geographical location to climate and climate change, making it possible to detect gradual as well as abrupt shifts in their thermal characteristics. Where these data sets include years with abnormal weather conditions, they are especially valuable, since this information may help us evaluate the likely response of the lakes to future extremes. Fritz (1996) postulated that lakes located in extreme habitats or near an ecotone or climatic boundary will respond most sensitively to climate change. In this respect, lakes located in the Alpine/Perialpine region, on the Atlantic coast, and at high latitudes are likely to respond most sensitively to the climatic changes summarised in Chapter 2.

Ecologically, even relatively small changes in the thermal characteristics of lakes – e.g. in their thermal stratification – can cause major shifts in phytoplankton, bacterioplankton and zooplankton populations as well as altering the rates of metabolic processes (e.g. Steinberg and Tille-Backhaus, 1990; Tulonen et al.,1994; Weyhenmeyer et al., 1999; Gerten and Adrian, 2000; Arvola et al., 2002; Jasser and Arvola, 2003). This is because organisms are often adapted to certain narrow temperature ranges and because their life-cycle strategies can be highly sensitive to variations in ambient water temperature (e.g. Chen and Folt, 1996).

6.2 The Lakes

This chapter gives a brief overview of our knowledge of the impact of the changing climate on the thermal characteristics of European lakes. Most of the results presented are based on statistical analyses of long-term records from the CLIME lakes and some additional lakes from Finland (Table 6.1). The 25 lakes included in the study are located in three very different climatic regions: Northern Europe, here comprising Estonia, Finland and Sweden; Western Europe, comprising Ireland and the UK; and Central Europe, comprising Austria, Germany and Switzerland. The response of these lakes to climate change depends on their mixing characteristics, which, in turn, are influenced by their location, landscape setting, size and topography. The lakes selected thus belong to very different mixing types. The Northern European lakes which experience winter ice cover are strictly dimictic; i.e., they undergo complete mixing twice a year – in autumn before ice-on and in spring after ice-off. They are thermally stratified both in summer (positively) and winter (negatively). The deep lakes in Central and Western Europe that are not ice-covered during winter are generally monomictic; i.e., they experience one period of complete mixing during the winter and a period of extended stratification during the summer. Some deep perialpine lakes can also experience incomplete mixing, and the thermal and chemical characteristics of the deep water in these lakes can differ quite substantially from year to year (Livingstone, 1993). Such lakes are usually referred to as facultatively monomictic (cf. Winter, 2004). Shallow lakes throughout the region, especially those with a large surface area, are usually polymictic; i.e., they mix frequently, or even continuously, throughout the year. If a shallow lake freezes, its mixing characteristics will of course change, as is the case in the

Table 6.1 Long-term trends and step changes in the summer and winter temperatures of selected European lakes. Statistically significant ($p<0.05$) long-term trends and step changes were analysed using a non-parametric Mann-Kendall test (MK; trends) and a distribution-free cumulative sum test (Cusum; step changes). MK: yes = statistically significant rising trend; yes (–) = statistically significant falling trend; no = no statistically significant trend. Cusum: year in which a statistically significant step change occurred; no = no statistically significant step change. ST = Surface-water temperature; BT = Bottom-water temperature

	Summer				Winter			
Country/Lake/Sampling period	ST		BT		ST		BT	
AUSTRIA (1961-2000)	MK	Cusum	MK	Cusum	MK	Cusum	MK	Cusum
Mondsee	yes	1991			no	no		
Hallstätter See	no	no			no	no		
Attersee	yes	1981			yes	no		
Traunsee	yes	no			no	no		
SWITZERLAND								
Lower Lake Zurich (1972-2006)	no	no	yes	1988	yes	1987	yes	1990
Upper Lake Zurich (1972-2006)	no	no	no	no	no	no	yes	1987
Greifensee (1960-2005)	no	no	no	no				
GERMANY (1979-2002)								
Müggelsee	yes	1991						
UK (1960-2000)								
Windermere (South Basin)	yes	no	no	1985	yes	1986	yes	1981
Windermere (North Basin)	yes	1988	no	no	yes	1988	yes	no
Esthwaite Water	yes	1988	no	no	no	no		
Blelham Tarn	yes	no	no	no				
IRELAND (1960-2000)								
Lough Feeagh	yes	no			no	no		
ESTONIA (1960-2004)								
Peipsi	no	no						
Võrtsjärv	no	no			no	1997		
FINLAND (1970-2005)								
Artjärven Pyhäjärvi					yes	1998	no	no
Lohjanjärvi 1 (Isoselkä Basin)	no	no	no	1993	no	1998	no	no
Lohjanjärvi 2 (Karjalohjanselkä Basin)					no	no		
Vanajavesi	no	no	no	no	no	1998	yes	1993
Säkylän Pyhäjärvi	no	no	no	no	yes	1988		
Päijänne 1 (Ristiselkä Basin)	no	no						
Päijänne 2 (Tehinselkä Basin)					no	1987		
Pääjärvi					no	1987		
SWEDEN								
Vättern (1979-2003)	no	1988			no	1988	no	no
Mälaren (Galten Basin) (1965-1995)	yes (–)	no			no	1988		

two large Estonian lakes (Võrtsjärv and Peipsi) covered by the CLIME project. In contrast, shallow, sheltered lakes, such as the many small lakes found in the boreal region, may be strictly dimictic, temporarily monomictic or meromictic (i.e., the deep water layers stagnate and do not mix).

6.3 The Statistical Analyses

The statistical methods used to analyse the data collated here include a non-parametric Mann-Kendall test for monotonic trends and a non-parametric, distribution-free, cumulative sum test (Cusum) for step changes (eWATER Catchment Modelling Toolkit, Catchment Hydrology, Trend C1.0.2). Linear regression analysis was used to calculate rates of change of temperature. All the statistical analyses were performed on the raw, measured data. Details of the measurements, such as the time periods on which the analyses were based, are given in the appropriate figures and tables.

6.4 Winter Water Temperatures

The long-term records analysed here indicate that during the past few decades there has been a weak, common, rising trend in winter water temperatures in lakes throughout Europe (Table 6.1). Figure 6.1 shows the year-to-year variations recorded in the surface temperature of 17 of these lakes. When the collated results were pooled by country (Table 6.2), the mean rate of increase of the winter surface temperature varied from 0.014 K year^{-1} in Austria to 0.026 K year^{-1} in the UK, and the bottom water temperature from <0.010 K year^{-1} in Finland to 0.034 K year^{-1} in the UK Statistically significant trends in both the winter surface-water and bottom-water temperatures were recorded in only two lakes; viz. Windermere (UK) and Lower Lake Zurich (Switzerland). The rate of increase of the winter surface-water temperature was 0.026 K year^{-1} for Windermere and 0.022 K year^{-1} for Lower Lake Zurich. Combining the results for the surface water temperatures of all northern lakes, a statistically significant increasing trend is found, although taking the lakes separately, the increasing trend was statistically significant in only two out of the ten lakes analysed. The evidence for a sustained rising trend is thus quite weak and the key point to note is the variability of the response observed in the different lakes.

When the step changes in the data sets were analysed, the Finnish and Swedish lakes behaved very coherently, indicating an upward jump in water temperature under the ice around 1987/1988 (Table 6.1). The timing of this jump was almost the same in Windermere and Lower Lake Zurich. These step changes in water temperature occurred in synchrony with the corresponding changes in winter air temperature.

In strictly dimictic, ice-covered lakes at high latitudes and/or high altitudes, warmer winters will tend to result in a shorter period of ice cover, with lakes freezing

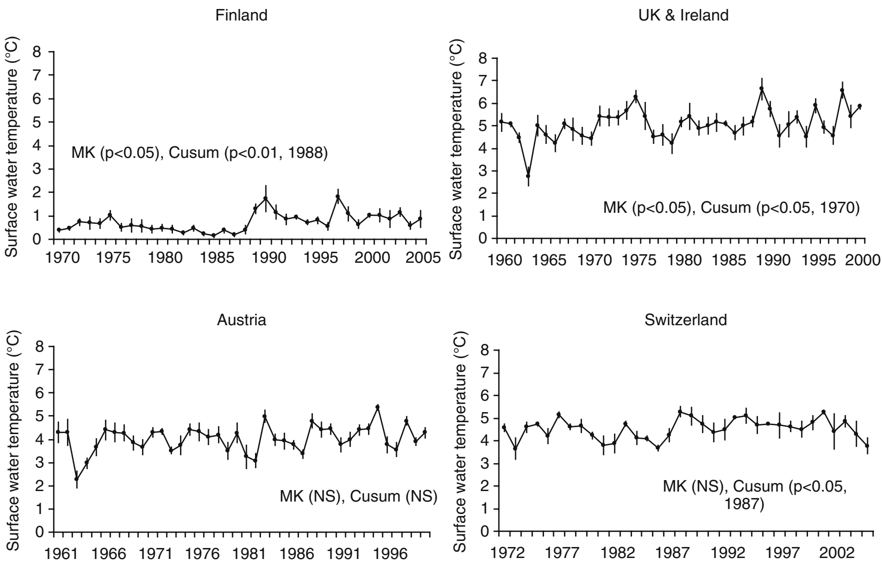

Fig. 6.1 Mean winter surface water temperature (±1 standard deviation) in eight lakes in Finland, three in the UK and Ireland, four in Austria and two in Switzerland during the last 30 years or longer. A non-parametric Mann-Kendall test (MK) was used to test for the existence of a monotonic trend, and distribution-free cumulative sum statistics (Cusum) to determine the likelihood of a probable step change in temperature. The relevant statistical significance for the existence of a monotonic trend and a step change are given, along with the year of the probable step change. NS = no statistically significant ($p < 0.05$) trend or step change. Winter is defined as January–March for the Austrian and Swiss lakes, as the middle of March for the Finnish lakes, and as December-February for the UK and Irish lakes. The definitions differ because of differences in the climatic conditions prevailing at the lakes. n = number of lakes (see Table 6.1). Note the different scales on the x-axes

later and thawing earlier than in colder winters (cf. Chapter 4). This will result in turnover being later in autumn and earlier in spring. Furthermore, the earlier thawing of snow and ice on the lakes can result in earlier occurrence of the phytoplankton spring bloom, as has been found, for example, in Lake Erken, Sweden (Weyhenmeyer et al., 1999). Later freezing, possibly preceded by a longer autumnal mixing period, may cool the entire water column in deep lakes more efficiently than is possible during shorter autumnal mixing, and may also result in better oxygenation of the hypolimnion.

However, in monomictic lakes, such as most of the large, low-altitude lakes in western and central Europe, warmer winters will result in a decrease in the frequency, intensity and penetration depth of mixing events. In lakes which always mix in winter because of continually high wind speeds, lake temperatures will increase at all depths, forced by the increasing equilibrium temperature at the air-water interface. Data from monomictic lakes in the UK imply that a long-term increase in both surface and bottom-water temperatures has indeed occurred (Table 6.2). In lakes which do not always mix completely each year, the frequency of occurrence

Table 6.2 Long-term increases or decreases in water temperatures in winter and summer in 19 European lakes (i.e. a subset of those listed in Table 6.1), pooled by country (countries with less than three lakes were omitted). Listed are the gradients of the relevant linear regressions (in K yr^{-1}). ST = surface-water temperature, BT = bottom-water temperature, N = number of lakes from each country. Min and max are the minimum and maximum gradients found in each of the countries.

		Winter ST			Winter BT		
Country	N	mean	min	max	mean	min	max
Finland	8	0.016	0.006	0.030	0.008	–0.005	0.021
UK	4	0.026	0.022	0.029	0.034	0.026	0.050
Austria	4	0.014	0.003	0.029			
Switzerland	3	0.022	0.002	0.046	0.011	0.002	0.016
Mean		0.019			0.018		
		Summer ST			Summer BT		
Country	N	mean	min	max	mean	min	max
Finland	8	0.038	–0.006	0.079	0.005	–0.019	0.045
UK	4	0.035	0.030	0.039	0.007	–0.002	0.010
Austria	4	0.043	0.017	0.054			
Switzerland	3	0.029	0.017	0.040	0.000	–0.010	0.008
Mean		0.036			0.003		

of events in which thermal stratification persists uninterruptedly from one summer to the next (although becoming weaker in winter) will increase. Because of the slow downward mixing of warmer water from the lower metalimnion into the hypolimnion by turbulent diffusion, such events result in a gradual increase in deep-water temperature, which at some point will reach a level high enough for complete mixing to occur regardless of the degree of severity or mildness of the winter (Livingstone, 1997). Such long-term gradual warming followed by rapid cooling results in an irregular sawtooth structure in the deep-water temperature time-series (Fig. 6.2). This phenomenon has been observed in many lakes, with the period of gradual increase varying from 2 or 3 years in some lakes to over 20 years in others (Livingstone, 1993).

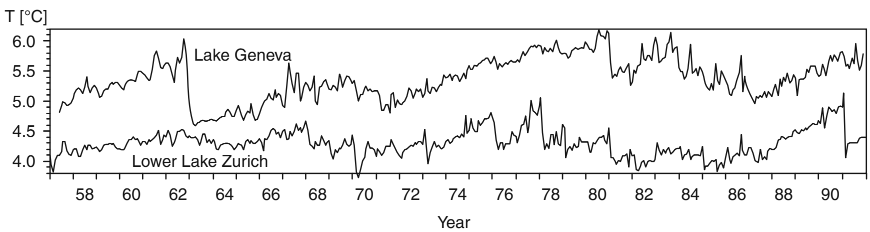

Fig. 6.2 Mean temperature below 100 m in Lake Geneva and Lower Lake Zurich from 1957 to 1991, showing the irregular 'sawtooth' temporal structure. Adapted from Livingstone (1993)

6.5 Summer Water Temperatures

The long-term records also indicate the existence of a rising trend in summer surface-water temperature in most of the selected lakes during the last few decades (Figs. 6.3 and 6.8, Tables 6.1 and 6.2). However, these trends are statistically significant in only nine of the 21 lakes investigated, with all nine being situated in either Western or Central Europe. In Northern Europe, no lakes show rising trends, and one lake even shows a falling trend (Table 6.1).

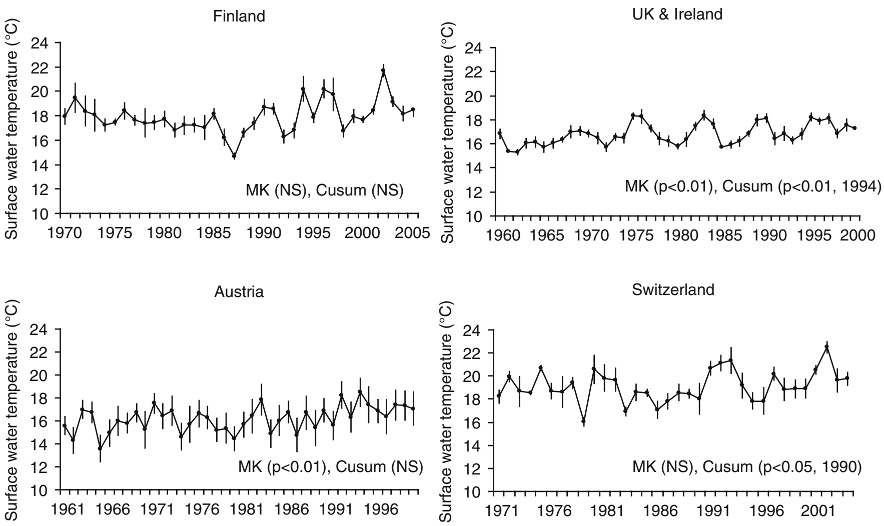

Fig. 6.3 As Fig. 6.1, except for summer. Summer is defined as June-August except for the Finnish lakes, for which it is defined as the middle of August

When these data sets were pooled by country (Table 6.2), the long-term mean rate of increase of summer lake surface water temperature varied from 0.029 K year^{-1} in Switzerland to 0.043 K year^{-1} in Austria, but the bottom-water temperatures exhibited essentially no change (< 0.007 K year^{-1}).

As in the case of winter air temperatures, the response of the lakes to increasing summer air temperatures also depends on lake type. In thermally stratified lakes, mixing characteristics are critically important. Figure 6.4A illustrates the strong year-to-year interdependence of the mean summer surface temperature of Esthwaite Water, a small lake located in the English Lake District, and the thermal stability of the water column. In this lake, the key factor influencing the summer surface temperature and the stability of the water column is the frequency and intensity of wind-mixing events, which, in turn, are related to the number of calm, anticyclonic days (Chapter 16, this volume). Figure 6.4B shows the relationship between the stability of the water column and the mean summer wind speed. Although the wind speeds in this hilly area are quite variable, the linear regression still accounts for more than 25% of the observed interannual variation.

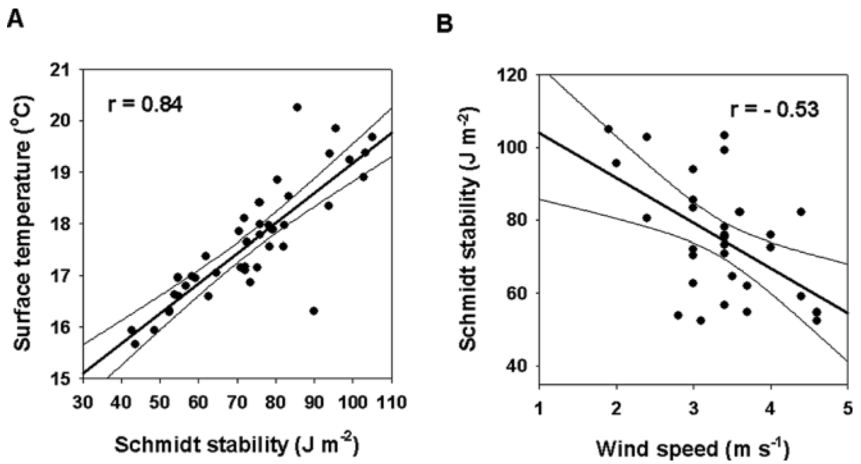

Fig. 6.4 (**A**) The relationship between the mean summer (July–August) surface-water temperature of Esthwaite Water and the thermal stability of the water column (calculated according to Schmidt, 1928). (**B**) The relationship between the mean summer stability of Esthwaite Water and the mean wind speed recorded at Ambleside

In strongly stratified lakes, this increased stability effectively shields the hypolimnion from the effects of the increasing air temperature. This effect is particularly pronounced in small, highly coloured lakes, where the epilimnion can be extremely shallow, sometimes less than 1 m. The difference in temperature between the epilimnion and the hypolimnion in these lakes can be very high (Fig. 6.5), a feature discussed in some detail by Arvola (1984) and Bowling and Salonen (1990). In such systems, the intensity of water colour can be an important regulator of thermal

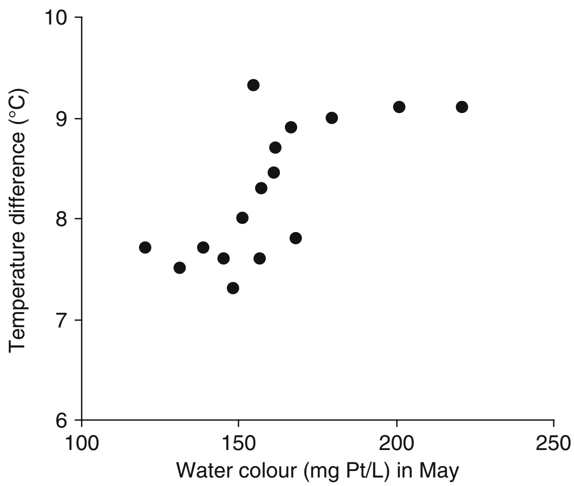

Fig. 6.5 The temperature difference (May–September mean, 1990–2005) between the epilimnion and the hypolimnion of Lake Valkea-Kotinen, Finland, as a function of the water colour in May, when thermal stratification first sets in

stratification and can also have a major effect on the underwater light climate and the ecology of the lake (Jones and Arvola, 1984).

Thus, in thermally stratified lakes, increasing summer air temperatures will only result in increasing hypolimnetic temperatures if winds are strong enough to result in mixing, often mediated by internal waves, between the metalimnion and the hypolimnion. In this case, increasing air temperatures together with increasing wind stress may also increase the total heat content of a stratified lake by deepening the epilimnion. Otherwise, the temperature of the hypolimnion will be primarily determined by the conditions that pertained during the previous spring turnover. In fact, the increased thermal stability that follows from increased summer air temperature can result in a lower hypolimnetic temperature, as shown by model studies (Hondzo and Stefan, 1993) and observations (Livingstone and Lotter, 1998; Tanentzap et al., 2008). In shallow, polymictic lakes, the entire water column responds to fluctuations in air temperature, so these lakes are likely to respond more directly to short-term variations in the weather.

In all cases, it is the thickness of the mixed layer, which can be the epilimnion or the whole water column, that determines how rapidly a lake responds to such fluctuations in air temperature. This is illustrated well by the behaviour of the surface temperature of three CLIME lakes in northern Europe (Võrtsjärv in Estonia, and Valkea-Kotinen and Pääjärvi in Finland) during a period of very hot weather in July 2003 (Fig. 6.6). In Võrtsjärv and Valkea-Kotinen, the diel fluctuations in surface water temperature were pronounced and very similar, whereas in Pääjärvi they were much less pronounced, although the surface water temperatures of all three lakes exhibited a similar longer-term variability. This is interesting because the Finnish

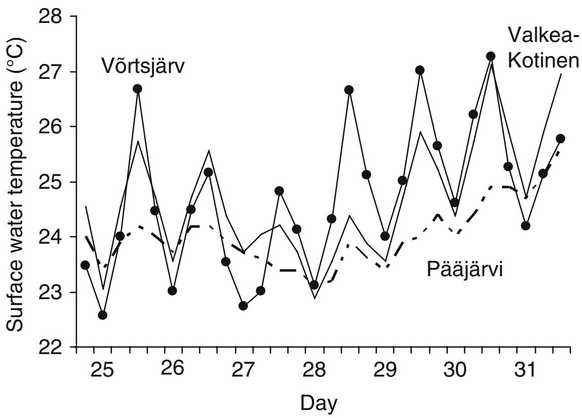

Fig. 6.6 The diel fluctuation of surface-water temperature in one lake in Estonia (Võrtsjärv) and two in Finland (Pääjärvi and Valkea-Kotinen) during warm weather from 25 to 31 July 2003 (during the last three days the maximum daily air temperature commonly exceeded 30°C in Estonia and southern Finland). The measurements were obtained using automatic temperature data loggers. The mean surface temperature of each lake during the period shown varied between 24.0°C (Pääjärvi) and 24.7°C (Võrtsjärv and Valkea-Kotinen)

lakes are located only 20 km apart, but Võrtsjärv lies 300 km to the south of the Finnish lakes. Võrtsjärv is a large, shallow (mean depth 2.8 m), polymictic lake, while Valkea-Kotinen and Pääjärvi are stratified humic lakes. Valkea-Kotinen has a thin epilimnion, but in Pääjärvi the epilimnion in summer is between 5 and 10 m deep, and is thus more resistant to rapid changes forced by varying air temperature.

6.6 The Timing of Thermal Stratification

Since air temperatures recorded in winter and spring have increased in recent years, the onset of spring turnover in most northern lakes has generally become earlier. This trend is typically easier to detect in the lakes that are now free of ice for a long period. However, data from Lake Erken, in Sweden, suggest that there has been no corresponding advance in the end of spring turnover (Weyhenmeyer et al., 1999), implying that the duration of spring mixing may also have increased. In Lake Erken, the duration of the summer stratification period has clearly increased, and the end of this period now occurs almost one month later than it did in the 1960s (Fig. 6.7). In contrast, small, sheltered, humic lakes may stratify immediately after ice-out if the weather is sunny, warm and calm. This kind of temporary 'spring meromixis' means that such lakes do not necessarily turn over completely in spring (Salonen et al., 1984), a situation which has been documented several times in Valkea-Kotinen (Finland) since 1990.

In the Western European lakes, the timing of the onset of stratification is more difficult to quantify, but significant long-term shifts have been detected in Windermere, the largest lake in the English Lake District (Thackeray et al., 2008). In Lower Lake Zurich, in Switzerland, a continual long-term increase in thermal stability resulted in an increase of about two weeks in the duration of the summer

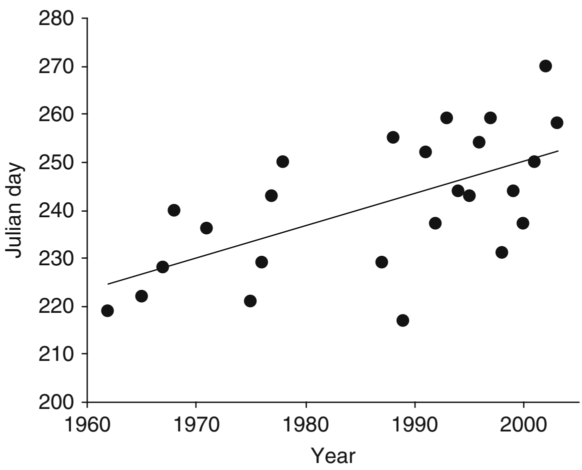

Fig. 6.7 The timing of the end of summer stratification in Lake Erken, Sweden, since the early 1960s. The linear trend ($r^2 = 0.378$) was statistically significant ($p < 0.01$)

stratification from the 1960s to the 1990s. Shifts in the timing of both the beginning and end of the stratification period contributed to this, although the shift in the timing of the end of the period was more consistent and less variable than the shift in the timing of the beginning (Livingstone, 2003). The observations in Lake Erken and Lower Lake Zurich support the results of modelling studies conducted to predict the effects of climate change on lakes (e.g., Hondzo and Stefan, 1993; Stefan et al., 1998; Peeters et al., 2002).

6.7 Discussion

In this chapter, we have compared water temperature records from a number of lakes and shown that there is now convincing evidence of a long-term increase in their surface temperature, which appears to have accelerated during the 1990s. These pan-European comparisons complement the results of more detailed regional studies such as those of Livingstone and Dokulil (2001), Livingstone (2003) and George et al. (2007b). The most rapid increases in both the winter and summer temperatures are those reported from lakes in the UK and Ireland (George et al., 2007a, 2007b). In central Europe, especially in the alpine and northern perialpine regions, the thermal behaviour of lakes is more variable. However, water temperatures in lakes of varying size, bathymetry and location can behave coherently to a certain extent there also, as has been observed in Swiss and Austrian lakes by Livingstone (1993, 1997), Livingstone and Dokulil (2001), and Dokulil et al. (2006). At the end of the 1980s, a consistent, abrupt increase in winter water temperatures occurred in lakes throughout Western and Northern Europe, which is perhaps somewhat surprising in view of the fact that the northern lakes are ice-covered in winter whereas the western lakes are not. This suggests that the causal mechanisms underlying the observed step changes may not be the same in these two regions, although in both cases they ultimately result from large-scale warming.

Different types of lakes, and lakes in different geographical and climatic regions, respond in different ways to the observed variations in the weather. The bar graphs in Fig. 6.8 summarise the results of the statistical tests reported in Table 6.1. The number of significant long-term trends is shown for the winter and summer time-series collated for the selected lakes in Northern, Western and Central Europe.

In Northern Europe, statistically significant rising trends were found only in a few lakes, which in fact is in good agreement with previous observations from North America (e.g., Magnuson et al., 2006). Similarly, Korhonen (2002) was not able to identify any common statistically significant trend in an analysis of the surface water temperatures of eight large Finnish lakes from 1961 to 2000. According to Korhonen (2002), surface water temperatures increased significantly only in the southernmost part of Saimaa, the largest lake in Finland, and in particular during late summer. The reason for this remains unclear, but it may be related to changes in the optical properties of the water as a consequence of anthropogenic loading, e.g. eutrophication (see Jones et al., 2005) and/or changes in shore vegetation, as Korhonen

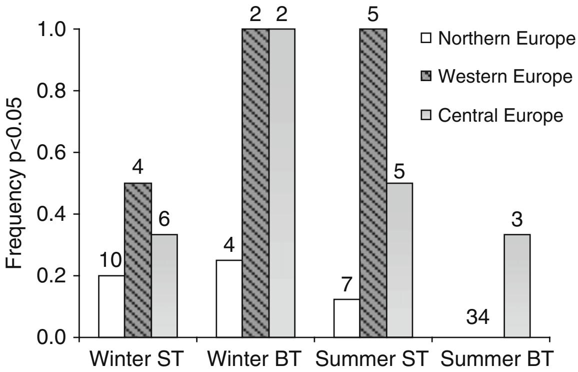

Fig. 6.8 Mean frequency of statistically significant long-term trends in water temperature in European lakes, determined using the Mann-Kendall test (for more details see Figs. 6.1 and 6.3, and Table 6.1). Abbreviations: ST = surface-water temperature, BT = bottom-water temperature. The number of lakes is given above each of the columns. In the case of summer BT the values for northern and western Europe are so close to zero that the bars are not visible

(2002) speculates. However, the data imply that summer water temperatures in the 1990s were above their long-term mean, which is consistent with the results from the Finnish lakes presented here. The main factor influencing the thermal characteristics of lakes in Northern Europe is the winter covering of ice. At the northernmost locations it is extremely rare for lakes not to freeze over during winter. During the exceptionally warm winter of 1988–1989, however, some lakes in central Sweden, e.g. Mälaren, did not freeze over. The mildness of this winter resulted in an increase in surface water temperature in seven of the 12 lakes in the Finnish and Swedish data set, although all the lakes in Finland were covered with ice. In Mälaren, this extreme situation resulted in an unusually early spring phytoplankton bloom with negative impacts on fishing and drinking-water supplies. The frequency of ice-free winters in southern Sweden has increased during the last 15–20 years, particularly in large lakes (Weyhenmeyer et al., 2005). In this regard, the largest lake in Europe, Lake Ladoga, in Russia, is likely to respond quite rapidly to future warming, since this lake is, on average, completely covered with ice only in February. As a result, in the future the behaviour of the lake may shift from dimictic to monomictic.

The CLIME lakes in Western Europe were all situated on the west coast and were consequently strongly influenced by the movement of weather systems across the Atlantic. These lakes consequently experienced the strongest shifts in their surface temperature in both summer and winter, but the increases in bottom-water temperature reported during summer were very much less. At some sites, the temperatures recorded in the hypolimnion may be negatively correlated with the air temperature, in situations very similar to those described by Livingstone and Lotter (1998), Dokulil et al. (2006) and Tanentzap et al. (2008). This is because the downward vertical transport of warm metalimnetic water into the hypolimnion is less efficient in hot summers than in cold summers due to the increased thermal stability

(Livingstone and Lotter, 1998; George et al., 2007b). Hondzo and Stefan (1993) observed the same phenomenon in their study of north-central US lakes. According to them, the overall lake stability in spring and summer will become greater as the climate warms, which is also in line with the observations of temporary spring meromixis in humic lakes. Climate-induced changes in the catchment and in the hydrology of the inflows may enhance the loading of allochthonous dissolved organic carbon (cf. Chapter 13) to a lake, which in turn may alter its optical properties by increasing light attenuation. This may produce stronger thermal stratification, as documented above, which can be important particularly in small boreal and arctic lakes. Very similar conclusions were reached by Fee et al. (1996) working on Canadian Shield lakes.

Most of the CLIME lakes in Central Europe were relatively deep and integrated the imposed climatic signal over much longer periods of time (see Chapter 17). In deep lakes with irregular patterns of circulation, such as Lake Geneva, unusually cold winters such as those experienced in 1963 and 1981 can result in very low thermal stability over a long period of time, leading to deeply penetrative mixing and cooling of the deep water (Livingstone, 1993; Dokulil et al., 2006). However, if deep lakes freeze over, as happens in the case of all deep lakes in Finland, most deep lakes in Sweden, and as happened in both Lower Lake Zurich and Windermere during the cold winter of 1963, the water column under the ice is essentially cut off from external sources of kinetic energy, thus inhibiting vertical mixing. The climate change scenarios summarised in Chapter 2 suggest that summer air temperatures in Central Europe could be 6–8°C higher by 2071–2100. Increases of this magnitude would have a profound effect on the lakes, and we can already gain some impression of the changes involved by reviewing their response to previous warm summers.

Extremely warm summers, like that of 1976 in Western Europe, 2002 in Northern Europe and 2003 in Northern and Central Europe, produce pronounced increases in the surface water temperatures and the thermal stability of lakes. In two Swiss lakes (Lower Lake Zurich and Greifensee), for instance, Jankowski et al. (2006) showed that the warm summer of 2003 resulted in high temperatures in the epilimnion and metalimnion, which, combined with hypolimnetic temperatures that were slightly lower than normal, gave rise to steep temperature gradients and increased thermal stability. Increases in surface water temperature and thermal stability were also observed in northwestern Ontario, Canada, during the 1970s and 1980s as a result of the occurrence of weather that was warmer and drier than normal (Schindler et al., 1990, 1996). Surface heating, especially when combined with reduced wind mixing, can also have consequences for the hypolimnion, as high thermal stability often results in oxygen depletion in the deep water in spite of lower-than-average water temperatures, a phenomenon that was also demonstrated in the two Swiss lakes by Jankowski et al. (2006). However, in some lakes mixing may bring more heat into the hypolimnion, which in turn can intensify the biodegradation of organic matter and cause oxygen depletion in the course of time. In Lake Mälaren, the unusually intense cyanobacterial bloom that occurred in the autumn of 2002 has been linked to high water temperatures in summer and autumn and to an increase in the transfer of phosphorus from the anoxic hypolimnion (Weyhenmeyer et al., 2004). Extreme weather events may sometimes have a long-term influence on the thermal conditions of lakes, as in the case of Lake Geneva in the winter of 1963, where the effect of deep, penetrative mixing on deep-water temperature lasted at least eight to nine years (Livingstone, 1993). In typically dimictic lakes, the effects of extreme events usually disappear during the next turnover event. The deep-water memory effect is therefore likely to last longer than one year only in lakes with incomplete annual turnover, such as meromictic lakes and facultatively monomictic deep lakes (Winter, 2004; Straile et al., 2003).

Sometimes extreme thermal conditions, even if they are of short duration, can produce severe chemical and biological responses in lakes (cf. Stefan et al., 1996). For instance, abnormally early freezing in autumn can leave the water column warmer than in a 'normal' year because it has less time to cool during mixing, which in turn may increase rates of mineralisation and oxygen consumption under ice, resulting finally in anoxia. Oxygen depletion can cause mass destruction of fish populations, as happened in winter 2002–2003 in southern Finland (Olin and Ruuhijärvi, 2005). In Lake Peipsi, in Estonia, high water temperatures and heavy algal blooms resulted in massive fish kills in 1959, 1972 and 2002. The risk of summer fish kills increases in hot summers because prolonged calm periods favour the development of bloom-forming cyanobacteria (Nõges, 2004). In larger lakes, water temperature may also have spatial and temporal gradients which can cause zooplankton development to occur at different rates in different areas of the lake, resulting in cascading effects on the food webs. This has been noted in Lake Constance, Germany (Straile, 2000) and Lake Washington, USA (Romare et al., 2005).

As the above examples show, year-to-year changes in the weather have a major effect on the thermal characteristics of lakes. Lakes behave as integrators of local weather and sometimes can even amplify weak meteorological signals, as has been documented, for instance, by George and Taylor (1995) and George (2006). The data summarised here demonstrate that, whilst the surface temperatures of lakes in Europe have tended to increase in recent decades, the pattern of change varies from lake to lake and region to region. These changes have important consequences for the ecology of the lakes and the effective management of water resources. In future, these changes may also have far-reaching socio-economic consequences. For example, during the last 10–15 years, people living in the southern part of Fennoscandia have already begun to respond to the changed winter conditions and to the late arrival and early disappearance of ice on many lakes. Traditional pastimes, like ice-fishing, have become increasingly difficult to pursue in southern Finland. One unfortunate consequence of poor ice conditions has been that a number of people have drowned after venturing on to thin ice. Another consequence has been that commercial fishing, involving the setting of gill nets and seine nets from the ice, has had to be abandoned at several sites.

Long-term temperature records of the kind analysed here are still surprisingly rare. Such information is, nevertheless, very useful and provides a very cost-effective way of detecting subtle patterns of change, quantifying the sensitivity of lakes to the changing climate, and providing the information required for the future management of these important resources.

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Chapter 7
Modelling the Impact of Climate Change on the Thermal Characteristics of Lakes

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7.1 Introduction

The projected changes in the climate (Chapter 2 this volume) will have a major effect on the thermal characteristics of lakes throughout Europe. A summary of the temperature changes observed in a number of CLIME lakes has been given in Chapter 6. Some of the ecological changes associated with these variations are also discussed in Chapters 18, 19 and 20. These results demonstrate that changes in the weather have a direct effect on the thermal characteristics of lakes and an indirect effect on the recycling of nutrients and the growth of phytoplankton. Thermally stratified lakes are particularly sensitive to changes in the weather since they integrate the effects of changes in the solar radiation, the air temperature and the wind speed (Chapter 17). There is now convincing evidence that the physical characteristics of many lakes in Europe have changed in recent years and that many of these variations are related to a systematic change in the weather (Weyhenmeyer et al., 1999; Livingstone and Dokulil, 2001; George et al., 2007).

Modelling studies play a key role in exploring the processes responsible for these changes since they can be used to test the sensitivity of the lakes to both the observed and projected changes in the climate. Over the years, a variety of one-dimensional models have been used to explore the potential impact of climate change on lakes. These include studies by Hondzo and Stefan (1993a); De Stasio et al. (1996) and Stefan et al. (1996) in North America and Frisk et al. (1997) and Peeters et al. (2002) in Europe. In thermally stratified lakes, these predict substantial increases in the epilimnetic temperatures, earlier onset of stratification and increases in the duration and strength of stratification. The projected changes in hypolimnetic temperatures are less consistent with decreases as well as increases being recorded in some stratified lakes. Most of the modelling work reported to date has been based on the outputs from relatively low resolution Global Climate Models (GCMs). Different models have often been used to support work in different regions (Peeters et al.,

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2002; Bell et al., 2006; Komatsu et al., 2006) and very little attention paid to the variation associated with their application to different types of lakes. In this chapter, we use an established lake modelling tool (PROBE) to explore the impact of the projected changes in the climate on the thermal characteristics of three lakes located in three very different climatic regions. The model simulations are driven by the outputs from the Hadley Centre Regional Climate Model (HadRM3p), and results presented for two contrasting greenhouse-gas emission scenarios. That the model is driven by high resolution RCM output and is used on three morphologically different lakes, allows an important question to be addressed, to whit, will the regional variation in future climate change or the morphological characteristics of a particular lake be the more important in determining the response of a lake to future changes in climate.

7.2 The Lake Model

7.2.1 The Development of the Lake Model

A one-dimensional lake model has been used to simulate the vertical temperature distribution in the three, different CLIME lakes: Lake Erken in Sweden, Esthwaite Water in the United Kingdom and Mondsee in Austria. As vertical gradients in lakes are larger than gradients in the horizontal, the short computation time and associated potential for multiple runs offered by a one-dimensional model is seen as being more advantageous for this work than any extension into further dimensions. Computation time is just a few minutes when simulating the temperature conditions experienced over several years in the lakes. The lake model used is based on a differential equation solver called PROBE (PROgram for Boundary layers in the Environment) which was developed by Urban Svensson (Svensson, 1978). It is written in Fortran and the code has been improved several times over the last 20 years. The present version is from 1998 (Svensson, 1998). A lake model based on this code was first developed in 1983 (Sahlberg, 1983) in order to investigate the effects of cooling on the vertical distribution of temperature in a Swedish lake. At this time, the model did not include any solar radiation or bottom water mixing terms but the calculated temperature profiles were in good agreement with the field measurements. The next published version of the model (Sahlberg, 1988) included solar radiation, heat exchange between water and sediment and a simple model that described the formation, growth and decay of lake ice. This was used to simulate the water temperature profiles under ice from the date of ice formation to the date of ice break-up.

The present version of the model has been improved by the inclusion of a bottom water mixing term, as the standard PROBE mixing model had proved unable to simulate realistic levels of mixing in the hypolimnion during periods of stratification. Additionally, through-flow of river water has also been included in order to simulate the seasonal variations in the water level and the internal dispersion of heat, nutrients and particulates. This version has already been successfully applied to a lake in northern Sweden (Sahlberg, 2003). Different versions of the lake model have also been used for lake studies in Finland, particularly for work on water temperature and ice formation (Elo and Vavrus, 2000; Elo, 2001). The same model has also been used for eutrophication studies in lakes (Malve et al., 1991) and to investigate how lakes respond to an assumed future climate change (Elo et al., 1998; Virta et al., 1996; Huttula et al., 1996).

7.2.2 The Lake Model Equations

The model is a finite-difference model, with the modelled lake being divided into multiple horizontal layers. Vertical mixing and temperature structure are derived from heat and momentum fluxes at the lake boundaries. Thus, the principal variables driving the model are wind speed, air temperature, cloud cover and relative humidity. A brief overview of the critical model equations is offered below.

The model solves a number of differential equations. The momentum equations are,

$$\frac{\partial \rho U}{\partial t} + W\frac{\partial \rho U}{\partial z} = -\frac{\partial P}{\partial x} + \frac{\partial}{\partial z}\left(\frac{\mu_{eff}}{\rho}\frac{\partial \rho U}{\partial z}\right) + f\rho V \tag{7.1}$$

$$\frac{\partial \rho V}{\partial t} + W\frac{\partial \rho V}{\partial z} = -\frac{\partial P}{\partial y} + \frac{\partial}{\partial z}\left(\frac{\mu_{eff}}{\rho}\frac{\partial \rho V}{\partial z}\right) - f\rho U \tag{7.2}$$

where t is the time variable, x and y horizontal space coordinates, z vertical space coordinate (measured upwards from the bottom of the lake), U and V horizontal velocities, W the vertical velocity (caused by in- and out-flows at different depths in the lake), P pressure, ρ density and f is the Coriolis term. The dynamical effective viscosity, μ_{eff}, is the sum of the turbulent dynamical viscosity, μ_{T}, and the laminar dynamical viscosity, μ. The heat equation is,

$$\frac{\partial(\rho c_p T)}{\partial t} + W\frac{\partial(\rho c_p T)}{\partial z} = \frac{\partial}{\partial z}\left(\frac{\mu_{eff}}{\rho\sigma_{eff}}\frac{\partial(\rho c_p T)}{\partial z}\right) + F_s^*(1-\eta)e^{-\beta_w(H-z)} + F_q \tag{7.3}$$

where T is the water temperature, c_{p} is the specific heat of water, σ_{eff} the effective Prandtl/Schmidt number which is the sum of the laminar and turbulent Prandtl/Schmidt numbers. $F_{\text{s}}{}^*$ is the part of the incoming short wave radiation that penetrates the water surface, η is the fraction of $F_{\text{s}}{}^*$ absorbed at the water surface, β_{w} is the extinction coefficient and H is the total depth. F_{q} is a source term due to the horizontal advection.

Mixing within PROBE is performed using a k-ε type model, whereby the dynamic eddy viscosity is calculated from the turbulent kinetic energy, k, and its dissipation rate, ε, by the Prandtl/Kolmogorov relation,

$$\mu_T = C_\mu \rho \frac{k^2}{\varepsilon} \tag{7.4}$$

where C_μ is an empirical constant. The turbulent kinetic energy is given by,

$$\frac{\partial k}{\partial t} + W\frac{\partial k}{\partial z} = \frac{\partial}{\partial z}\left(\frac{\mu_{eff}}{\rho\sigma_k}\frac{\partial k}{\partial z}\right) + P_s + P_b - \varepsilon \tag{7.5}$$

where σ_k is the Prandtl/Schmidt number for k. P_s and P_b represent shear and buoyant production of turbulent kinetic energy. The dissipation rate of turbulent kinetic energy becomes

$$\frac{\partial \varepsilon}{\partial t} + W\frac{\partial \varepsilon}{\partial z} = \frac{\partial}{\partial z}\left(\frac{\mu_{eff}}{\rho\sigma_\varepsilon}\frac{\partial \varepsilon}{\partial z}\right) + \frac{\varepsilon}{k}\left(C_{1\varepsilon}P_s + C_{3\varepsilon}P_b - C_{2\varepsilon}\varepsilon\right) \tag{7.6}$$

where σ_ε is the Prandtl/Schmidt number for ε. $C_{1\varepsilon}$, $C_{3\varepsilon}$ and$C_{2\varepsilon}$ are empirical constants. A full description of the derivation of all turbulence equations and empirical coefficients is found in Rodi (1980) and Svensson (1998).

The surface boundary conditions for the momentum equations are defined as,

$$\frac{\mu_{eff}}{\rho}\frac{\partial(\rho U)}{\partial z} = \tau_{ax} \tag{7.7}$$

$$\frac{\mu_{eff}}{\rho}\frac{\partial(\rho V)}{\partial z} = \tau_{ay} \tag{7.8}$$

where $\tau_{ax} = \rho_a C_D U_a W_a$ and $\tau_{ay} = \rho_a C_D V_a W_a$ are the wind stresses. Wind speeds in the along lake (x) and across lake (y) directions are described by U_a and V_a while $W_a = ({U_a}^2 + {V_a}^2)^{1/2}$ is the total wind speed. C_D is the drag coefficient and ρ_a is the air density. The drag coefficient, and the heat transfer coefficients, C_H and C_E (below), depend on atmospheric stability, as described by Launiainen (1995). At the lower boundary the condition of zero velocity is used.

7.2.3 The Bottom Turbulence Component

The standard version of the k-ε model describes the turbulence in the surface layer down to the thermocline. Below this layer, however, it often fails to predict a realistic turbulence field as the turbulence in deeper layers is usually not generated by the large-scale shear that the model resolves. In these layers it is believed that the turbulence is primarily generated by the shear produced by internal waves. Often the bottom layer kinematic diffusivity, ν_b, (Gargett, 1984; Stigebrandt, 1987) is formulated as

$$\nu_b = \alpha_N N^{-1} \tag{7.9}$$

where α_N is an empirical constant and N is the buoyancy frequency defined as,

$$N^2 = -\frac{g}{\rho_0}\frac{\partial \rho}{\partial z} \tag{7.10}$$

where g is the acceleration of gravity and ρ_o is the water density. This formulation was developed and tested for oceanographic situations and hence might not be valid for lakes and reservoirs. In a study made by Hondzo and Stefan (1993b) in four Minnesota lakes it was found that α_N could be formulated as a function of lake surface area and that a general expression for ν_b was,

$$\nu_b = 8.17 \times 10^{-8}(A_s)^{0.56}(N^2)^{-0.43} \tag{7.11}$$

where A_s is surface area (in km^2). The maximum value of ν_b is obtained under very weak stratification, defined as $N^2 = 7.0 \times 10^{-5}$ (Riley and Stefan, 1987). The new PROBE model uses the formulation by Hondzo and Stefan (1993b) for ν_b which is then added to the effective dynamic viscosity, μ_{eff}, calculated by the k-ε model.

7.2.4 The Heat Flux Component

For the heat equation the surface boundary condition may be formulated as,

$$\frac{\mu_{eff}}{\rho\sigma_{eff}}\frac{\partial \rho c_p T}{\partial z} = F_n \tag{7.12}$$

where F_n is the net heat exchange at the water surface. It consists of the sum of four different heat fluxes

$$F_n = F_h + F_e + F_{nl} + \eta F_s^* \tag{7.13}$$

where F_h, F_e, F_{nl} and $F_s{}^*$ are the sensible and latent heat fluxes, net long-wave radiation and short-wave radiation penetrating the water surface. A summary of the heat fluxes, which follow from Greenfell and Maykut (1977) and Maykut (1982), follows below and a full description can be found in Sahlberg (1988).

The insolation towards a surface is taken from Bodin (1979),

$$F_s = \varepsilon_t S_0 \cos Z(f_t - f_a)(1 - f_c) \tag{7.14}$$

where ε_t is the atmospheric turbidity, S_0 is the solar radiation outside the atmosphere, Z is the zenith angle of the sun, f_1, f_a and f_c are transmission, absorption and cloud functions, respectively, dependent upon optical path length and total cloud cover. S_0 is not a true constant but varies during the year due to the elliptical path of the earth's movement around the sun. In addition to the surface heat flux there is a sediment heat flux, F_{sed}, and a net heat flux due to the difference in heat between the in- and out-flowing water, F_{qin} and F_{qout}. A schematic of the heat fluxes is shown in Fig. 7.1. Note that, radiative long-wave, sensible and latent heat fluxes between water and air are inhibited during periods of ice-cover.

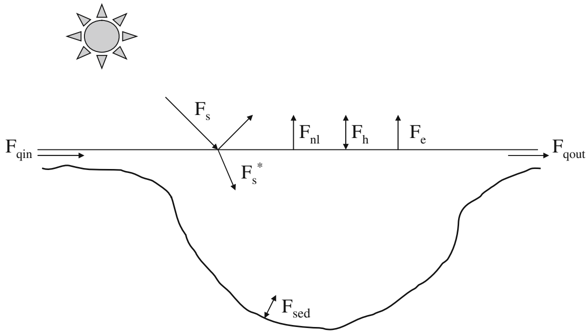

Fig. 7.1 The heat fluxes controlling the heat content of an ice-free lake. F_s is the short wave radiation and $F_s{}^*$ is the part penetrating the water surface. F_h and F_e are the sensible and latent heat fluxes. F_{nl} is the net long-wave radiation. F_{sed} is the sediment heat flux and F_{qin} and F_{qout} are the heat fluxes due to the through flow

The surface heat flux formulations are fully described by Omstedt and Axell (2003). The sensible and latent heat fluxes are formulated as,

$$F_h = \rho_a c_{pa} C_H W_a (T_s - T_a) \tag{7.15}$$

$$F_e = \rho_a L_e C_E W_a (q_s - q_a) \tag{7.16}$$

where c_{pa} is the heat capacity of air, L_{e} the latent heat of evaporation, C_{H} and C_{E} are the bulk exchange coefficients for sensible and latent heat respectively, T_{s} is surface water temperature, T_{a} air temperature. q_{s} is the specific humidity of air at the sea surface which is assumed to equal the saturation value at the temperature T_{s}, calculated as,

$$q_s = \frac{0.622 R_s}{P_a} \exp\left(\frac{q_1 T_s}{T_s + 273.15 - q_2}\right) \tag{7.17}$$

where $R_{\text{s}} = 611$, q_1 and q_2 are constants and P_{a} is the air pressure. The specific humidity of air, at the reference level, q_{a}, is accordingly calculated as,

$$q_a = \frac{0.622 R_s R_h}{P_a} \exp\left(\frac{q_1 T_a}{T_a + 273.15 - q_2}\right) \tag{7.18}$$

where R_{h} is the relative humidity ($0 \le R_{\text{h}} \le 1$). The formulations of bulk exchange coefficients are taken from Rutgersson et al. (2001) where neutral case values are from DeCosmo et al. (1996) and the stability dependence follows Launiainen (1995). Formulation of the net long-wave radiation, F_{nl}, is given by the difference between the upward and downward long-wave radiation, following Bodin (1979),

$$F_{nl} = \varepsilon_s \sigma_s (T_s + 273.15)^4 - \sigma_s (T_a + 273.15)^4 (b_1 + b_2 \sqrt{e_a})(1 + b_3 n_c^2) \tag{7.19}$$

where ε_{s} is the water surface emissivity, σ_{s} is the Stefan-Boltzmann constant and b_1, b_2 and b_3 are constants. Water vapor pressure, e_{a}, is related to q_{a} by,

$$e_a = \frac{P_a}{0.622} q_a \tag{7.20}$$

7.2.5 The Ice Model Component

When a lake is covered with ice, a number of changes take place to the modelled thermodynamics. Firstly, the momentum flux from the atmosphere to the water surface is cut off leading to zero horizontal velocities. Secondly, the only surface heat flux which penetrates the ice is the solar radiation, and even this heating is greatly reduced and depends on the albedo, thickness and extinction coefficient of the ice. The extent of penetration is further diminished by any snow that accumulates on the ice. Under such conditions, the heating caused by the movement of water into and out of the lake becomes much more important together with the exchange of heat between the deep water and the bottom sediments. There is also a small transfer of heat between the water and the ice.

In the simulations described here, the ice component was based on the simple model developed by Sahlberg (1988). This model was primarily designed to simulate the annual variation in the heat content of a frozen lake and included simple routines for estimating ice formation, growth and decay. The model assumes that ice formation occurs when the surface water temperature becomes super-cooled. The initial ice formation may be rather complex, with ice forming one day and breaking up the next. This behaviour is parameterised in the following way. If a lake is ice covered and the ice thickness is less than 0.1 m the ice cover will break up and melt if the daily mean value of the wind velocity is greater than 4 ms^{-1}. When the ice thickness is larger then 0.1 m the ice growth is calculated using the well known degree-day method (Bengtsson and Eneris, 1977). The melting of the ice in spring is taken from Ashton (1983) and is based on daily mean air temperature. This ice model has been tested on a small lake in Sweden, Lake Tulebo, with good agreement between measured and calculated ice formation, growth and decay (Sahlberg, 1988). In CLIME, some considerable time was devoted to developing and testing the new ice-model described in Chapter 5. This model includes representations of the slush and snow-ice layers in the ice sheet but only became available in the closing stages of the project.

7.3 The Sites Used for the PROBE Simulations

The three lakes selected for the PROBE simulations have been the subject of particularly intensive scientific study for more than 30 years. One lake was included from each climatic region and the type selected was broadly representative of the type commonly found in that region. Thus the Swedish lake was wide and relatively shallow and the Austrian lake narrow and very deep. Table 7.1 summarises the physical characteristics of the three lakes arranged in descending order of latitude.

Lake Erken is the largest of the three lakes modelled and covers an area of about 24 km^2. It is located near to the Baltic coast, roughly 70 km to the north-east of Stockholm (59.8°N, 18.6°E). Although generally shallow, with an average depth of 9 m, it contains a moderately deep trough dipping down to 21 m. This part of

Table 7.1 The physical characteristics of Lake Erken, Esthwaite Water and Mondsee

Lake	Area (km^2)	Max depth (m)	Mean depth (m)	Retention time (Years)	Latitude (°N)	Ice-cover	Trophic status	Altitude (m)
Lake Erken	23.7	21	9	6.5	60	Yearly	Mesotrophic	11
Esthwaite Water	1	15.5	6.5	0.3	54	Rarely	Eutrophic	26
Mondsee	14.2	68	36	1.7	48	Rarely	Mesotrophic	481

Sweden is relatively flat; the lake is near sea level with an altitude of only 11 m. The lake is surrounded by a small forest. Typical rainfall over the area is 550 mm per year. The lake has a particularly long residence time of 6.5 years. Winters are cold and the surface waters of the lake habitually freeze, usually for 3 or 4 months, with maximum ice thicknesses of approximately 0.3–0.4 m. The lake is regarded as being mesotrophic to eutrophic.

Esthwaite Water is a small (1 km^2) eutrophic lake in the Lake District, a mountainous area in the north-west of England (54.5°N, 3°W). It is morphologically complex, containing three sub-basins, the deepest of which is 15.5 m deep, although the average depth of the lake is only just over 6 m. The surrounding terrain is also complex with hills and mountains rising to a maximum height of a little under 1,000 m. As the area is close to the coast where the prevailing westerly winds traverse the Irish Sea, rainfall is frequent and is of order 1,500 mm per year. The average annual residence time for Esthwaite Water is about 90 days, although somewhat longer in the summer months and shorter during the winter. At one time, this site was covered with ice for several weeks in the year but very little ice has been observed on the lake in the last ten years.

Mondsee is the southernmost of the three modelled lakes. It is located in the northern, peri-Alpine region of Austria about 250 km west of Vienna (47.8°N, 13.4°E). It is a deep glacial lake slightly smaller than Lake Erken in surface area (14 km^2) but having substantially deeper maximum and average depths of 68 and 36 m, respectively. The region is mountainous, with the associated rainfall of almost 1,400 mm per year ensuring that lake water is renewed more quickly than in the less voluminous Lake Erken, the annual residence time being 1.7 years. Mondsee, at 481 m, is by far the highest of the lakes studied, but, nevertheless, has only sufficient elevation to freeze over occasionally. It is typically mesotrophic.

7.4 The Climate Scenarios Used to Drive the Lake Models

The physical lake models were driven by the outputs from two Regional Climate Models, the HadAM3 model developed by the Hadley Centre in the UK and the ECHAM model produced by the Max Planck Institute in Germany. The two models were perturbed by two contrasting greenhouse gas emission scenarios, the so-called

Table 7.2 The projected changes in the annually averaged meteorological variables for the grid boxes associated with the three lakes. The first values are based on the results from the Hadley Centre simulation with the A2 scenario (HCA2) and the results in parentheses are from the Hadley Centre simulation with the B2 scenario (HCB2)

Lake/variable	Air temperature (°C)	Wind speed (ms^{-1})	Cloud Cover (%)
Lake Erken	3.7 (2.4)	−0.1 (0.0)	−2 (0)
Esthwaite Water	2.4 (1.5)	−0.1 (−0.1)	−4 (−3)
Mondsee	4.1 (2.6)	−0.1 (−0.1)	−5 (−2)

A2 and B2 scenarios produced by the IPCC (2001). Both the models and scenarios are discussed at length in Chapter 2. Results from using the generated future scenarios from the two different models are compared in Persson et al. (2005). Since the two climate models produced very similar physical results, this Chapter only describes the outputs produced by the Hadley Centre model. Table 7.2 shows the major changes projected by this model as yearly averages. The meteorological variables listed are the air temperature, the wind speed and cloud cover. The values are based on the differences between the ‘control’ runs with the Hadley Centre model (HCCL) and the values projected by the same model for the A2 and B2 scenarios (HCA2 and HCB2).

As might be expected, the most pronounced change was that projected for the air temperature and the largest differences were those associated with the more extreme A2 scenario. There was no change in the projected wind speed and very little change in the cloud cover. This is, in part, because only yearly averages are shown here. When the seasonal cycle is considered there is more variation, with the largest deviations being projected for cloud cover during the summer. Further details of these seasonal cycles and the differences observed between the three regions can be found in Chapter 2.

The changes in the meteorological variables influence the flux of heat between the atmosphere and the lake in different ways. For example, an increase in cloud cover decreases the down-welling solar radiation but increases the downward long-wave radiation. On the other hand, an increase in wind speed will increase the magnitude of both the sensible and latent heat flux. The direction of these fluxes can be either upward or downward, albeit on average they are upward fluxes. The increase in air temperature will decrease the loss of sensible and latent heat from the lakes and also increase the downward long-wave radiation. The net result will be for the water temperature to increase. As it does so, the turbulent fluxes will decrease until a new equilibrium, with both warmer air and warmer water, is established.

The most striking aspect of these simulations is the marked variation in the air temperature projections for the different regions. This primarily reflects their geographic position with the most pronounced increase being suggested for the lake situated in the most ‘continental’ location away from the moderating effect of the Atlantic Ocean. The difference between the increases projected for Esthwaite Water and Mondsee is almost as great as the absolute increase projected for the lake in the

UK. The dominance of the air temperature as a heat budget parameter implies that the regional differences in the projected values of this variable will have the decisive effect on the future behaviour of the lakes in the three regions.

7.5 Model Validation

PROBE has previously been used successfully to model the thermal and mixing characteristics of a number of European lakes. Here, we present some example validations for Lake Erken in Sweden and Esthwaite Water in the English Lake District. Field measurements at these sites have been particularly intense and supplemented by high-resolution meteorological measurements. In this study, we used these measurements to validate the model for periods that included all four seasons. Both surface water temperatures and deep water temperatures are compared, the exact depth being dependent upon the depth at which the observations were taken. The years chosen for the validation coincide with the periods when the most complete temperature records were available. In Lake Erken, this year was 1992 when the lake was covered with ice for the first two months of the year, thawed in mid-March and stratified a few months later in early May. In summer, the maximum difference between the surface and the bottom water temperatures was about 10°C (Fig. 7.2A). Overturn took place in the late summer and the lake then cooled until ice-out later in the year. These features were well captured by the model but there was some disparity between the observed and predicted bottom water temperatures in late summer. In Esthwaite Water, these records covered the period between February 1997 and April 1998. No significant ice was recorded on Esthwaite Water in 1998 and the lake was already stratified by the end of April. Here, the pattern of summer heating was more complicated than that observed at Lake Erken and stratification lasted well into the autumn. The model fit for Esthwaite Water was also very good with the most pronounced deviations being recorded in the deep water in early summer. In a previous publication, Persson et al. (2005) used thirty-year averages of the seasonal variation in the water temperature to simulate the effects of the projected changes in the climate on Erken, Esthwaite Water and Mondsee. Results from these simulations showed there was a good agreement between the modelled and measured temperatures. Overall, the closeness of fit of modelled and observed data confirms the suitability of the PROBE model for simulating these lakes and provides a robust way of exploring their response to future changes in the climate.

7.6 The Temperature Projections for 2071–2100

The results of the PROBE modelling showed that changes in the climate, on the scale projected, would have a major effect on the thermal characteristics of the selected lakes. Figure 7.3a–c compares the seasonal variation in the surface and bottom water temperatures of the lakes with those simulated using the PROBE and Hadley

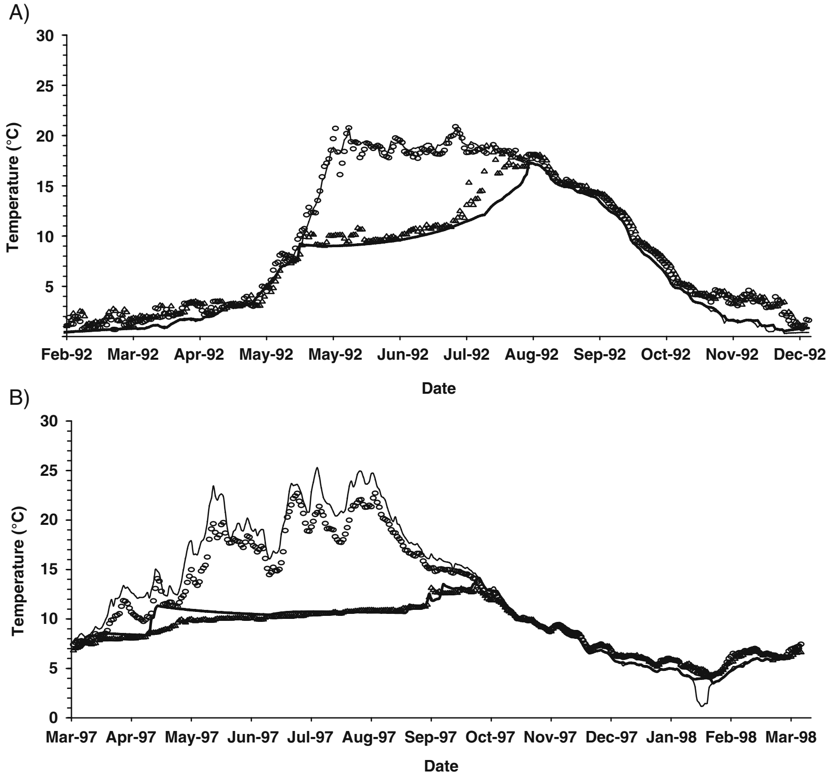

Fig. 7.2 The observed and modelled variation in the surface and bottom temperature of: (**A**) Lake Erken in Sweden and (**B**) Esthwaite Water in the UK. *Thin* and *thick lines* show modelled surface and bottom temperatures, and circles and triangles the observed surface and bottom temperatures

Centre model under present day (control) and future (warm world) conditions. The projections are for the period between 2071 and 2100 and the values shown are for the A2 (HCA2) and B2 (HCB2) emission scenarios. In most cases, the lakes show a substantial increase in both their surface and bottom temperatures but the magnitude of this increase varies from lake to lake and season to season. The only exception was the slight cooling of the bottom waters in Mondsee under the HCB2 scenario. This illustrates the complicated nature of hypolimnetic warming in lakes which are influenced both by the intensity of stratification and by the mixing effects of the internal seiche. Further examples of this differential warming and cooling are given in Chapter 6 which documents the long-term trends observed in a number of European lakes. The warming associated with the HCA2 scenario is about twice as large as that associated with the HCB2 scenario and the between site differences in the predicted summer temperatures reflect these differences. Not surprisingly, the most pronounced increases were those predicted for the surface waters. The predicted changes in the bottom temperatures were very much lower with those in

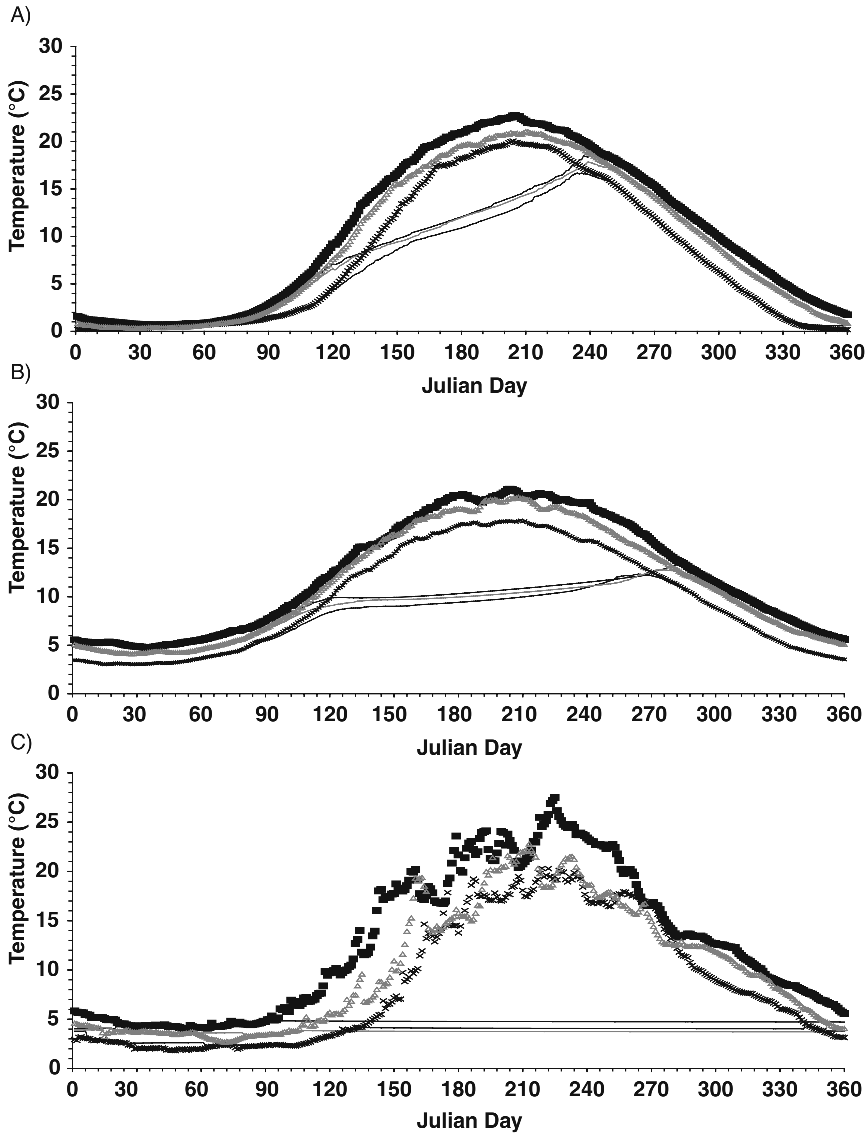

Fig. 7.3 The seasonal variation in the predicted surface and bottom temperature of: (**A**) Lake Erken; (**B**) Esthwaite Water and (**C**) Mondsee using the Hadley Centre model and the A2 and B2 scenarios. Crosses, grey triangles and large solid squares show the surface temperature for Control, B2 and A2 scenarios and *thick*, *grey medium* and *thin lines* show the bottom temperature for Control, B2 and A2 scenarios (modified from Persson et al., 2005)

Table 7.3 The predicted increase in the surface and bottom water temperatures of the three lakes. Results are presented for the summer and winter averages using the differences between the 'control' and 'warm world' simulations with the Hadley Centre Model. The first values listed are for the A2 scenario and the values in parentheses are for the B2 scenario (modified from Persson et al., 2005)

Lake/variable	Surface (summer) (°C)	Surface (winter) (°C)	Bottom (summer) (°C)	Bottom (winter) (°C)
Lake Erken	2.8 (1.4)	1.1 (0.5)	1.1 (0.8)	0.7 (0.3)
Esthwaite Water	2.9 (2.0)	2.1 (1.4)	0.8 (0.5)	2.0 (1.3)
Mondsee	5.4 (1.8)	2.5 (1.4)	0.7 (1.3)	1.6 (1.3)

Mondsee remaining low throughout the year. Table 7.3 summarises the predicted changes in the top and bottom temperatures of the three lakes generated by the three model simulations. The summer values for the two lakes are the averages for June, July and August. The winter values are averages for December−February for Esthwaite Water and Mondsee and December−March for Lake Erken.

An important feature of all the simulations was the change in the timing and intensity of thermal stratification. Some indication of these variations can be gained by comparing the curves for the surface and bottom water temperatures in Fig. 7.3a−c. Table 7.4 summarises the predicted change in the onset and the duration of stratification for the three lakes. The first values shown are the differences generated by the HCA2 scenario and the values in parentheses are those associated with the HCB2 scenario. Also shown is the difference between the projected increase in the surface and bottom temperatures for each lake which can be used as a general measure of the change in the intensity of stratification. In qualitative terms, the lakes responded in a very similar way with stratification starting earlier, lasting longer and increasing in intensity. Quantitatively, though, there were some obvious differences between the lakes. The most pronounced change in both the duration and intensity of stratification was that predicted for Mondsee. Here, the maximum increase in the duration generated by the HCA2 scenario was 63 days and there was a 4.7°C increase in the temperature difference between the surface and bottom

Table 7.4 The predicted change in the onset of stratification, the duration of stratification and the top minus bottom temperature in the three lakes. The differences are those predicted by the Hadley Centre Model. The change in the top minus bottom temperature is calculated as the average change in surface temperature minus the average change in bottom temperature for the period of stable stratification. The first values listed are for the A2 scenario and the values in parentheses are for the B2 scenario (modified from Persson et al., 2005).

Lake/variable	Change in onset (days)	Change in duration (days)	Change in top minus bottom temperature (°C)
Lake Erken	−15 (−8)	21 (16)	1.7 (0.6)
Esthwaite Water	−9 (−5)	24 (16)	2.1 (1.5)
Mondsee	−45 (−14)	63 (14)	4.7 (0.5)

water. It should be noted, of course, that the magnitude of change depended on the selected emission scenario. The qualitative similarity of the results, though, suggests that the projections are robust and provide a reasonable measure of the variability associated with the A2 and B2 projections.

7.7 Discussion

In CLIME, we used an established physical model (PROBE) to explore the potential impact of future changes in the weather on the thermal characteristics of three contrasting lakes. Although these lakes were located in areas exposed to very different climates, the validations showed that the model was able to provide a reasonable representation of the seasonal variations in temperature. CLIME was one of the first projects to make use of the climate projections generated by the latest generation of Regional Climate Models. The RCM used to drive the simulations summarised here had a spatial resolution of ca. 80 km and included projections for both the A2 and B2 emission scenarios produced by the IPCC. So what do the predicted changes in the temperature and physical stability of the selected lakes tell us about their sensitivity to the projected changes in the climate? The first point to note is the qualitative similarity of the responses predicted for the three lakes. All become warmer, stratify earlier and have longer and more intense periods of thermal stratification. Very similar changes in the timing and intensity of thermal stratification have been reported from lakes in North America (King et al., 1997) and very similar projections have been produced by other modelling studies (e.g. Hondzo and Stefan, 1993a; De Stasio et al., 1996).

The difference in the projected responses of the lakes can partly be explained by their geographic location and partly by their physical characteristics. Lake Erken is a large, relatively shallow lake whose physical, chemical and biological characteristics are strongly influenced by the duration of the ice-free period (see Chapters 8 and 18 this volume). The heating processes in an ice-covered lake are very different to those in a lake without ice-cover. The model predicts that the extra heat from the projected increase in the air temperature would be used in the hibernal months to melt the ice in the lake, rather than to warm the water. The lake ice-model used in PROBE is much simpler than that described in Chapter 5. Nevertheless, the results presented by Persson et al. (2005) suggest that there will be a dramatic reduction in the number of days when ice occurs on the lake, changing to an average of 16 days with the B2 scenario and an average of only 7 days with the A2 scenario. The temperature increases projected for the ice-free period at Erken are not particularly large, but the model results suggest that there could be a 16−21 day increase in the duration of thermal stratification. Esthwaite Water is a small lake located in a sheltered valley and is strongly influenced by year-to-year variations in the intensity of wind mixing (see Chapters 6 and 19). Here, the projected increases in the summer surface temperature are more pronounced but the increased stability of the water limits the amount of heat transferred into deep water. Mondsee is a large, deep lake situated in a mountainous region of Austria. As such, it might be expected to be less responsive to changing climatic conditions but our projections demonstrate that the opposite is true. Here, the increases in the water temperature, the intensity of stratification and the duration of stratification were far more pronounced than those predicted for the other sites. This difference can, almost entirely, be explained by the very large air temperature/cloud cover changes projected for this area by the scenarios described in Chapter 2. In the most extreme scenario, the summer surface temperature increased by 5 °C but the bottom temperatures remained low due to the increased stability of the summer water column.

The number of lakes covered by these model simulations is too small to draw any conclusions regarding the ameliorating effect of basin topography. It is clear, however, that changes in the intensity of stratification and the rate at which heat is transferred into deep water can have a major effect on the climatic response of a lake. In their study of the effects of the North Atlantic Oscillation on the seasonal dynamics of lakes in Germany, Gerten and Adrian (2001) showed that differences in the morphometry of the individual basins had a major effect on the vertical transfer of heat and the temperatures measured in deep water. Similarly, under the extremely warm conditions experienced in Europe during the summer of 2003, it was observed that, whilst Lake Zurich and Greifensee in Switzerland both set records with their epilimnetic temperatures in the summer, their hypolimnetic temperatures remained relatively low (Jankowski et al., 2006). Such observations are consistent with the model results of Hondzo and Stefan (1993a) who showed a reduction in hypolimnetic temperatures for some stratified North American lakes under a future climate scenario. They suggested that whilst hypolimnetic temperatures would rise in shallow lakes they would fall, by more than a degree, in deep lakes. Furthermore, for lakes of medium depth they predicted that a rise or a fall in the hypolimnetic temperature would be related to the size of the lake and the timing of stratification. The small lakes that stratify early in the year would experience the most pronounced reductions in their hypolimnetic temperatures. It seems likely that the complications associated with thermal stratification make future changes to their hypolimnetic temperatures much more difficult to predict than their epilimnetic temperatures. This would, in turn, lead to uncertainties about their future heat storage capacity and to the magnitude of changes to the strength of stratification.

In general, it appears that the location of a lake is probably the more important determinant in how it will respond to changes in climate. This is partly because of the strong regional variation in the predicted changes in climate, so that atmospheric forcing will simply change by a larger amount in some regions than in others. In part, it is also connected to some factors being of much greater relevance in some locations than in others; so, for example, extent of ice-cover is extremely important in more northern and eastern lakes. Morphology, though, still has an influence, with lake area and depth shaping how much stratification will change in lakes and, in particular, constraining the extent to which hypolimnetic waters will be altered by a changing climate. As the number of steps in the modelling process increases, from future atmospheric conditions to resulting lake temperatures and on to phytoplankton responses, uncertainty cascades down to the final results. There is therefore, now a need for modelling sensitivity studies on lake responses to individual atmospheric parameters, to examine systematically the effects of a changing climate, such as performed by Elliott et al. (2006), and for these to be extended to a whole suite of climate change scenarios, so that the integrated effect of individual parameter changes can be understood.

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Chapter 8
The Impact of the Changing Climate on the Supply and Re-Cycling of Phosphorus

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8.1 Introduction

For more than half a century, phosphorus (P) has been a major focus of limnological research. Early studies by Rodhe (1948) identified P as a key factor limiting the growth of algae. Recently, Istvánovics (2008) has reviewed the cycling of phosphorus in lakes and its role in eutrophication. Historical increases in P loading have been observed in lakes distributed throughout Europe. In urban areas, the main factor responsible for the increase has been the influx of P from municipal and industrial point sources (Forsberg, 1994). In rural areas, the increase in loading has principally been due to changes in land-use and the increased use of artificial fertilizers (Forsberg, 1987). Over the years, water companies and local administrations have successfully reduced the quantities of P reaching lakes from point sources but reducing the load attributable to diffuse sources has proved more difficult. Another factor influencing the seasonal availability of P to lakes and reservoirs is the internal recycling of the element from both shallow and deep sediments. During the 1980s research on the role of internal P loading from lake sediments developed rapidly and new knowledge was gained about factors regulating this process in both shallow and deep lakes (Boström et al., 1982; Cullen and Forsberg, 1988). Mobilisation as well as transport processes were identified and the classical Einsele-Mortimer theory (Einsele, 1938; Mortimer, 1941; Mortimer, 1942) that explained the reduction of iron and the dissolution of phosphate under anaerobic conditions was expanded to include the microbial breakdown of organic phosphorus as a significant factor. More recently, increasing attention has been paid to the effects that year-to-year variations in the climate have on both the supply and the internal recycling of phosphorus.

Catchments act as filters for phosphorus (Hillbricht-Ilkowska and Pieczynska, 1993) and can behave as sinks, attenuators or sources depending on the spatial and temporal distribution of phosphorus as well as its speciation. Year-to-year varia-

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tions in the air temperature and precipitation are of considerable importance and can affect both the retention of phosphorus in the catchment and its transport into the lakes. The climatic factors influencing the internal recycling of phosphorus are even more complex. They include the effect of wind-mixing on the vertical transfer of the nutrient and the effect of light on the proportion assimilated by the growth of phytoplankton. Here, the physical characteristics of individual lakes are of paramount importance. Lakes with a short retention time are typically more sensitive than lakes with a long retention time whilst the internal dynamics of phosphorus is very different in isothermal and thermally stratified lakes (Kamp-Nielsen, 1975; Boström et al., 1982; Pettersson, 1998; Søndergaard et al., 2003). The anticipated impacts of changes in the climate on the supply of phosphorus and the growth of phytoplankton are thus of growing concern to water managers (Lathrop et al., 1998; Schindler, 2001).

In Chapter 6, Pierson et al. use the outputs from two Regional Climate Models to drive a catchment model that simulates the impact of the projected changes in the climate on the supply of phosphorus in a number of different lakes. The catchment model used is the Generalized Watershed Loading Function (GWLF) described by Scheiderman et al. in Chapter 3. The climate projections are those described by Samuelsson in Chapter 2 and are based on the combined outputs of the RCAO and HadRM3p Regional Climate Models. In this chapter, we use a 'Case Study' approach to explore the effects of year-to-year variations in the weather on the supply and recycling of phosphorus in four lakes that are physically very different. Particular attention is paid to their response to extreme weather events and their sensitivity to both long-term and short-term changes in the weather. Further information on the dynamics and trophic structure of the lakes can be found in George (2002); Pettersson et al. (2003); Pierson et al. (2003); Nõges et al. (2003) and Weyhenmeyer et al. (2004).

8.2 The Choice of Sites and the Rationale for the Analyses

Long-term records of phosphorus concentrations in lakes are surprisingly rare and the analyses of many existing time-series are complicated by anthropogenic changes in the surrounding catchment. One way of minimizing the effects associated with a slow, progressive, change in the concentration of phosphorus is to de-trend the time-series to highlight the impact of the inter-annual variations in the weather. This approach was used by George (2000a) to demonstrate the impact of changes in the weather on the winter concentration of phosphorus in Windermere and has since been applied to other lakes in the English Lake District (George et al., 2004). Such techniques can, however, only be applied in situations where the lakes have been sampled at frequent intervals over a relatively long period of time. Only a small proportion of the lakes included in the CLIME project had phosphorus records that were sufficiently detailed for systematic analysis. Table 8.1 lists the four sites used for the case studies reported here. The sites are arranged in descending order of latitude and cover a range of sizes with different mixing characteristics.

Table 8.1 The general characteristics of the sites used in the three Case Studies

Site	Latitude	Surface area (km^2)	Residence time (days)	Maximum depth (m)	Thermal characteristics	Trophic status
Erken	59°25'N 18°15'E	24	2500	21	Dimictic	Meso-eutrophic
Võrtsjärv	58°15'N 26°00'E	270	350	6	Polymictic	Eutrophic
Esthwaite Water	54°21'N 2°59'W	1.0	90	14	Monomictic	Eutrophic
Blelham Tarn	54°23'N 2°58 W	0.1	40	11	Monomictic	Eutrophic

The first Case Study is based on Lake Erken, a moderately deep, dimictic, meso-eutrophic lake located in the south-east of Sweden. It has a surface area of 24 km^2, a volume of 213.5 10^6 m^3, a mean depth of 9 m and a maximum depth of 21 m. The summer stratification typically lasts from mid-June to mid-September (Pettersson and Grust, 2002). The lake is usually frozen for at least 4 months of the year, from mid-December to mid-April, and the estimated residence time is about 7 years (Weyhenmeyer et al., 1997). The lake has no inlets affected by point sources but several of the inlet streams are polluted with low concentrations of nutrients from diffuse inputs (Pettersson, 1985). The mean total phosphorus concentration during winter is about 24 $\mu g\ l^{-1}$ and the yearly mean concentrations of total nitrogen and chlorophyll *a* are 620 $\mu g\ l^{-1}$ and 3.7 $\mu g\ l^{-1}$, respectively. The extended residence time of the lake means that short-term changes in the composition of water reaching the lake has little effect on the inter-annual variations in the concentration of phosphorus. However, the internal processes that regulate the re-cycling of nutrients from the sediments have a profound effect on the seasonal dynamics of phosphate and the productivity of the lake (Istvánovics et al., 1993; Pettersson, 2001; Pettersson and Grust, 2002).

The second Case Study is based on Lake Võrtsjärv, a large, shallow, polymictic lake located in Central Estonia which has a surface area of 270 km^2 at average water level. It has a volume of 750 · 10^6 m^3, a mean depth of 2.8 m and a maximum depth of 6 m. Natural water level fluctuations can change the lake area by a factor of 1.4, the mean depth by a factor of 2.5 and the volume by a factor of 3.5. The lake is usually covered with ice from mid-November until mid-April and the average residence time of the lake is about one year. During the ice-free period, the Secchi depth is commonly less than 1 m, mainly due to the resuspension of sediments. Lake Võrtsjärv is highly eutrophic and is characterized by a mean annual total nitrogen concentration of 1.6 mg l^{-1}, a total phosphorus concentration of 54 $\mu g\ l^{-1}$ and chlorophyll *a* concentration of 24 $\mu g\ l^{-1}$ (Haberman et al., 1998). Agricultural activities and point source loadings from the towns of Valga (population 14,000) and Viljandi (population 21,000) provide the main source of these nutrients.

The third Case Study is based on data collated for two small, productive, lakes in the English Lake District. Esthwaite Water has a surface area of 1.01 km^2, a mean depth of 6.4 m and a retention time of about 90 days. Blelham Tarn has a surface area of 0.10 km^2, a mean depth of 6.8 m and a retention time of about 42 days. The two lakes are usually only covered with ice for a few days in the year and the period of stable stratification typically extends from early June until the end of September. The winter concentration of dissolved reactive phosphorus (phosphate) in the two lakes has always been high, but summer concentrations are close to the limits of detection. In this Case Study, we demonstrate the effects that year-to-year variations in the weather have on both the supply and internal re-cycling of phosphate. The main factor influencing the supply of phosphate is the quantity leached from the surrounding catchment. The concentrations measured in the lake are, however, regulated by biological uptake and a variety of re-cycling mechanisms. In the examples presented here, we consider both the effects of winter rain on the external supply of phosphate and the effects of the summer wind on its internal re-cycling.

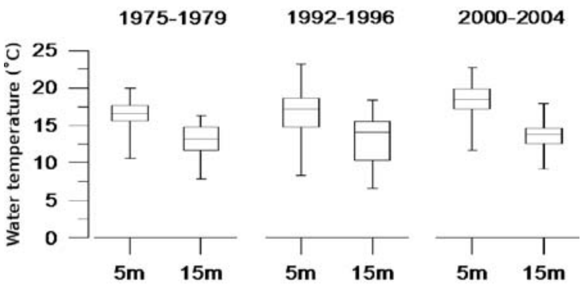
8.3 The Impact of the Observed Variations in the Weather on the Supply and Recycling of Phosphorus

A number of studies have already shown that the lakes used for these Case Studies are very sensitive to changes in the weather. The northern sites are strongly influenced by changes in the freeze-thaw dates of the lakes (Weyhenmeyer et al., 1999; Blenckner et al., 2004; Blenckner et al., 2006), and there has been a recent extension in the length of the ice-free season. In contrast, the two western sites are more sensitive to the effects of heavy winter rain and the effect of the wind on the entrainment of nutrients from deep water (George, 2002; George et al., 2007). In the sections that follow, we pay particular attention to the factors that influence the internal dynamics of phosphorus in these lakes and quantify their relative sensitivity to the observed variations in the weather.

8.3.1 Case Study 1: Lake Erken (Sweden)

The first measurements of water chemistry in Lake Erken were recorded in 1933. This early sampling was, however, somewhat irregular since the measurements were only designed to meet the needs of a particular project. The first regular sampling began in the 1970s and continued with little change until the late 1980s. In 1993, this sampling was intensified with samples collected at weekly intervals during the ice free period (April–October) and at monthly intervals for the remainder of the year. Since then, water samples have been collected at a station located near the

Fig. 8.1 The long-term change in the summer epilimnetic (5 m) and hypolimnetic (15 m) water temperatures in Lake Erken



deepest part of the lake. Water samples are normally collected at 2 m intervals in the water column and bulked to produce an integrated, sample that is proportional to the volume of water between some defined depths (Blomqvist, 1995). In the 1970s and 1980s the depths selected were 1, 5, 10, 15 and 19 m. During the period of stratification, the top three samples were usually in the epilimnion and the bottom two in the hypolimnion.

One of the most important factors influencing the year-to-year variations in the concentrations of phosphate in Lake Erken is the variation in the epilimnetic and hypolimnetic temperature. Figure 8.1 summarizes these variations as a series of box-whisker plots. The variability has been calculated for the stratified period and the values shown are those recorded at a depth of 5 m in the epilimnion and 15 m in the hypolimnion. The results demonstrate that there has been a small but consistent increase in the maximum temperatures and a marked increase in the variability of these measurements in the 1990s. Figure 8.2 shows the long-term change in the phosphate concentrations recorded in Lake Erken in the spring and during the winter. 'spring' has been defined as the first period of complete mixing and 'winter'

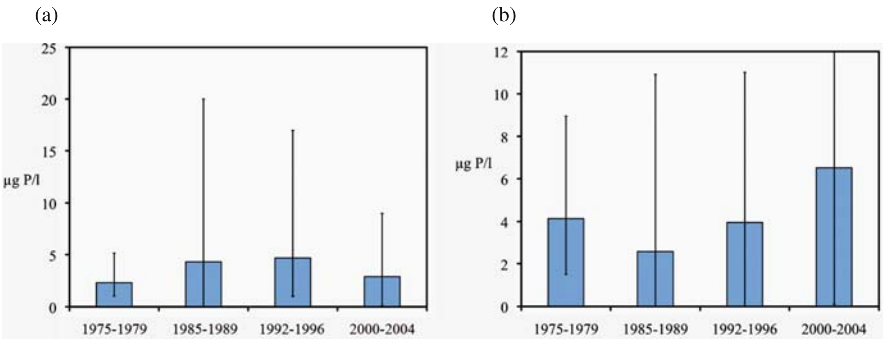


Fig. 8.2 The long-term change in the average phosphate concentrations recorded in Lake Erken in (**a**) the spring and (**b**) the winter. The histograms show the four year averages and the lines the maximum concentrations

as the period when the lake is covered with ice. The results demonstrate that there has been very little change in the spring concentrations (Fig. 8.2a), a feature that reflects the relatively constant load of nutrients from the catchment and the efficient uptake of phosphate by the growth of phytoplankton. In contrast, there has been a marked increase in the winter concentrations (Fig. 8.2b) with very high concentrations being recorded between 2000 and 2004. In Lake Erken, the concentration of phosphate in the open water is at its lowest at the time of the spring phytoplankton bloom and typically remains low throughout the summer. The increased concentrations now being recorded during the winter are primarily a reflection of the increased concentrations that accumulate in the hypolimnion during the summer. The concentration of phosphate present in the hypolimnion at the end of summer is influenced by a number of physical and biological factors. The most important physical factors are the timing of stratification, the stability of the thermocline and the timing and intensity of wind-induced mixing. From a biological point of view, the key factors are the underwater light regime, the seasonal succession of different species of phytoplankton and the length of the growing season. The changes in the weather experienced in the last decade have had a significant effect on these processes and given rise to a disproportionate increase in both the hypolimnetic concentration of phosphorus and the concentrations measured in the lake after the autumn overturn. Figure 8.3a, b compare the phosphate concentrations measured in the hypolimnion and epilimnion of Erken in four successive five year periods. Despite some considerable inter-annual variation, it is clear that there has been a marked increase in the concentrations measured in both the hypolimnion and the epilimnion in the period between 2000 and 2004. The main reason for this sudden increase was the extension of the ice-free period, the length of the growing season and the number of warm summer days. Blenckner et al. (2006) have described the consequent increase in the stability of the water column and the associated depletion of oxygen. A model simulation that demonstrates the future course of this effect is described in Chapter 15.

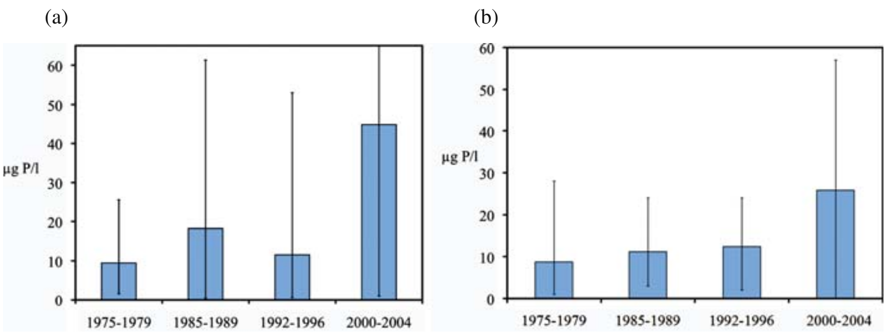


Fig. 8.3 The long-term change in the average phosphate concentrations recorded in Lake Erken during the summer in (**a**) the hypolimnion and (**b**) the epilimnion. The bars show the four year averages and the lines the maximum concentrations

8.3.2 Case Study 2: Lake Võrtsjärv (Estonia)

Monthly measurements of dissolved reactive phosphorus (phosphate) in Lake Võrtsjärv began in 1968 and were supplemented by measurements of total phosphorus (TP) in 1983. Chemical analyses were performed on depth integrated water samples using standard photometric analysis and the methods described by Grasshoff et al. (1983). Phosphate was measured by the molybdate blue method using ascorbic acid as the reductant. For TP determination, organic compounds were mineralized into phosphate, using persulphate.

The results show that the maximum loadings of phosphorus to Lake Võrtsjärv usually occur in winters that are either very wet or extremely cold. For example, the rainy years 1981 and 1998 were characterised by greatly increased inputs of phosphate. Similarly very high phosphate loadings were recorded in 1986 and 1987 when the winters were very dry and cold. Very cold winters are usually followed by sudden floods when large volumes of nutrient rich water from the upper layers of the soil enter the rivers and the lake. In Lake Võrtsjärv, the historical variations in the productivity of the system complicate the task of analyzing these weather-related effects. Analyses using de-trended data (Nõges et al., 2007) nevertheless suggest that year-to-year variations in the weather have had a significant effect on the annual flux of phosphate. When the phosphate time-series were de-trended to minimize the effects associated with the increased use of fertilizers, there was a significant positive correlation ($r = 0.6$, $p < 0.01$) between the annual loading of phosphate and the observed inter-annual variation in rainfall.

The other climatic factor influencing the flux of phosphate in Lake Võrtsjärv was the periodic wind-induced resuspension of the deeper layers of the bottom sediments. The phosphate concentrations in the pore water of the surficial sediments are relatively low (typically < 10 ug l^{-1}), since they have a high Fe/P ratio (26–30) and are frequently aerated by resuspension. However, when the water level is low, severe storms, such as the one recorded in September 1996, can entrain substantial quanitities of the deeper, anoxic sediments (Nõges and Kisand, 1999). Estimates suggest that this internal re-cycling has amajor effect on the annual phosphorus budget (Huttula and Nõges, 1998). In one 'extreme event' recorded in 1996, measurements showed that the quantity of nutrients released from the sediments on one stormy day was greater than the total that entered the lake throughthe inflows over the whole year (Nõges and Kisand, 1999).

In a shallow lake, like Lake Võrtsjärv, less severe changes in the water level can also have a significant effect on the measured concentration of phosphate. These effects are particularly pronounced in the spring when there are large year-to-year variations in the timing of snow melt. Figure 8.4a shows the seasonal variation in the concentration of soluble reactive phosphorus in the lake between 1968 and 2004. The highest concentrations are recorded between January and April with the most pronounced inter-annual variations recorded in April. Figure 8.4b shows the extent to which the water level variations recorded in April influenced the concentration of phosphate. In this analysis, we have considered only the data acquired between 1991 and 1994, a period when the quantity of fertilizer used in the catchment

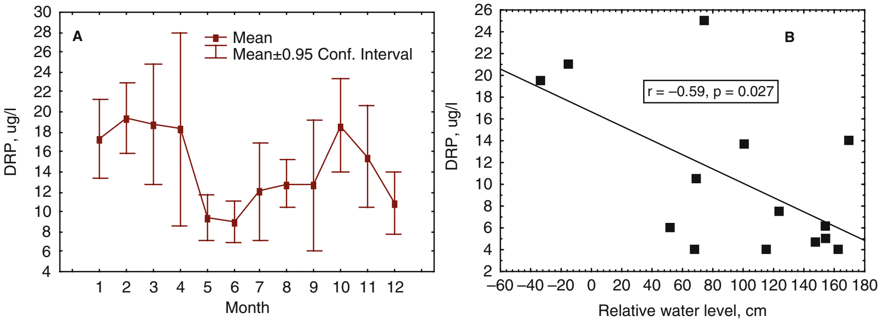


Fig. 8.4 (**A**) Seasonal cycle of dissolved reactive phosphorus (DRP) concentration (phosphate) in Lake Võrtsjärv in 1968–2004. (**B**) The relationship between the concentration of phosphate and the water level in recorded April (1991–2004 data).

had declined following the cessation of soviet-style agriculture. Once the effect of these loadings had been reduced, significant negative correlation could be detected between the concentration of phosphate and the variations in lake level. The water level fluctuations in Lake Võrtsjärv are principally governed by the weather conditions experienced during the winter. When the winters are cold and dry, the spring water levels are significantly lower and more phosphate is transferred into the water column by the resuspension of bottom sediments. Nõges et al. (2003) have shown that much of this inter-annual variation can be explained by changes in the status of the North Atlantic Oscillation. Positive values of the North Atlantic Oscillation Index (NAOw) are thus associated with mild winters, high water levels and lower spring concentrations of phosphate. Negative values are associated with drier winters, low water levels and an increased flux of phosphate from the bottom sediments (Fig. 8.5).

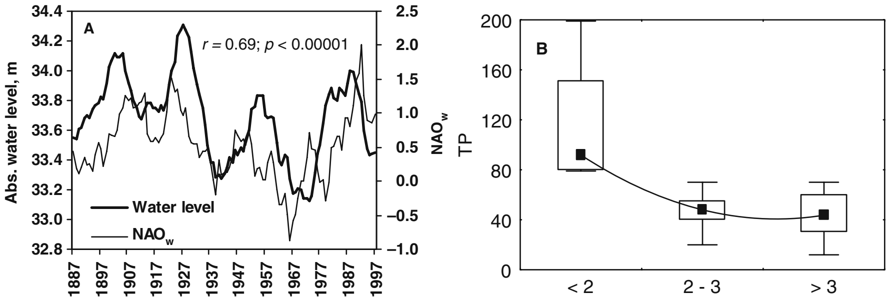


Fig. 8.5 (**A**) The year-to-year variations in the water level in Võrtsjärv in relation to the NAO. (**B**) The influence of the observed variations in water level on the total phosphorus concentration (μg P/l) (redrawn from Nõges et al., 2003)

8.3.3 Case Study 3: Esthwaite Water and Blelham Tarn (UK)

The data analyzed here, cover the thirty year period between January 1 1970 and December 31 2000. Daily measurements of the wind speed and the rainfall were recorded at a site situated about 4 km from the two lakes. The vertical variations in the water temperature were recorded using a thermistor attached to an oxygen electrode (1966–1984) and a profiler manufactured byYellow Springs Instruments (1984–1989). Samples of water for chemical and biological analysis were collected at a central station in each lake using a 5 m long plastic tube of the type described by Lund and Talling (1957). In the 1970s these samples were collected at weekly intervals but fortnightly sampling was introduced during the winter of 1979 and extended to the full year in 1992. Phosphate concentrations were measured by a modification of the solvent extraction method described by Stephens (1963).

8.3.3.1 The Seasonal Variations in the Concentration of Phosphate

The seasonal variations in the concentration of phosphate followed the same pattern in the two lakes with high concentrations recorded during the winter and low concentrations in summer. Figure 8.6 shows the average concentrations recorded in Esthwaite Water between January 1970 and December 2000. In these lakes, the seasonal dynamics of phosphate can best be understood by dividing the year into a series of ten week periods (quintiles) (George et al., 2000). Thus,the first quintile (Q1) covers the period when the rate of biological uptake is low, the second (Q2) coincides with the spring growth of diatoms, the third (Q3) the onset of thermal stratification, the fourth (Q4) the period of stable stratification and the fifth (Q5) the autumn overturn. In the following sections, we use the data acquired during the Q1 period as a measure of the external supply of phosphorus, and that acquired during the Q5 period as a measure of its internal recycling.

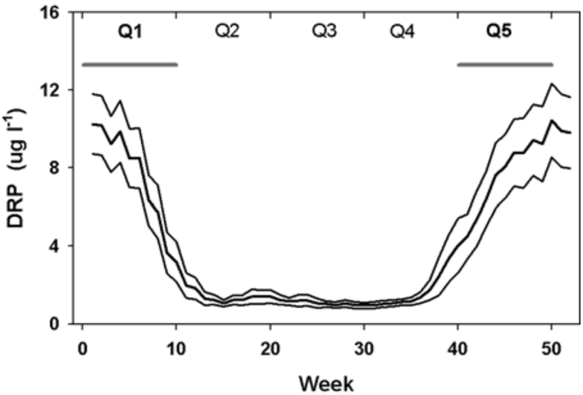


Fig. 8.6 The seasonal variation in the concentration of dissolved reactive phosphorus (phosphate) measured in Esthwaite Water between 1970 and 2000. The *bold line* shows the average concentration and the *lighter lines* the 95% confidence intervals. The *horizontal bars* show the two 'quintile' periods used in the analyses

8.3.3.2 The Long-Term Change in the Trophic Status of the Lakes

The productivity of most lakes in the English Lake District has increased progressively over the years. A number of recent studies (e.g. George, 2000a and George et al., 2004) have shown that these anthropogenic effects can be minimized by de-trending the raw time-series to highlight the year-to-year variations in the flux of phosphorus. Figure 8.7a, b shows the long-term change in the winter (Q1) concentration of phosphorus measured in Blelham Tarn and Esthwaite Water between 1970 and 2000. Sutcliffe et al. (1982) have shown that these winter measurements provide a convenient surrogate for the external loading since the proportion assimilated by

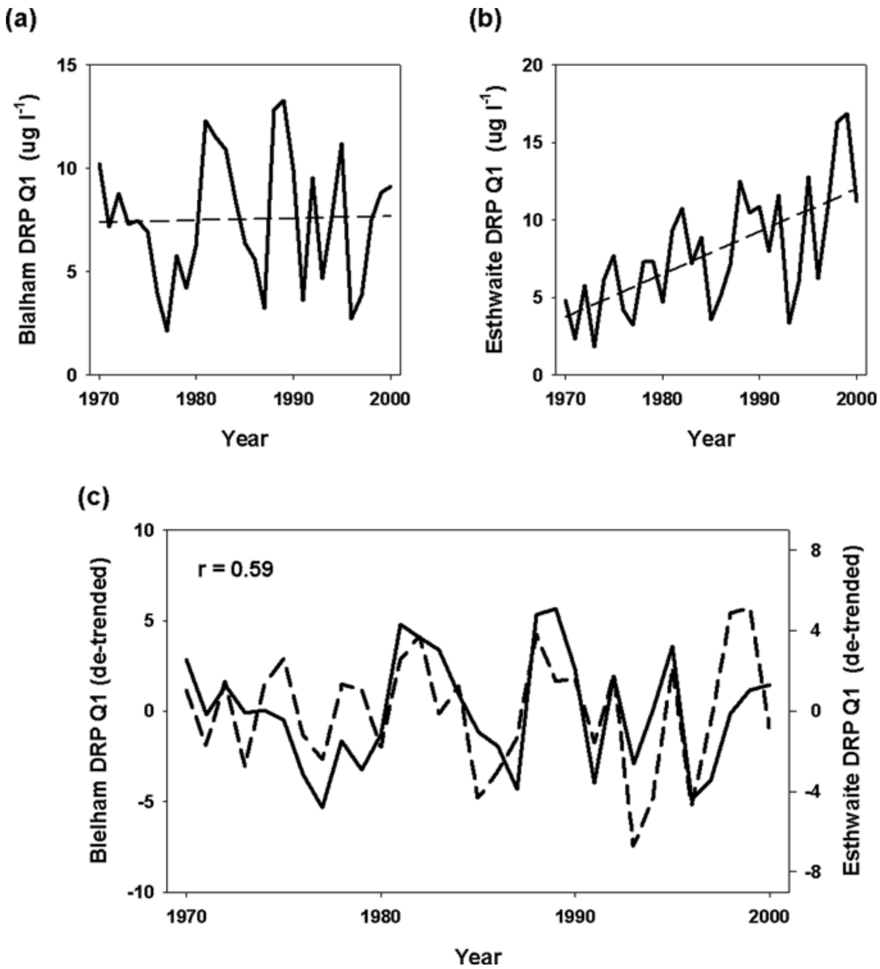


Fig. 8.7 The year-to-year variations in the winter (Q1) phosphate concentrations in (**a**) Belham Tarn and (**b**) Esthwaite Water. (**c**) The year-to-year variations in the de-trended concentration of phosphate (μg P/l). The *broken line* shows the results for Blelham Tarn

the biota is then very low. In Blelham Tarn (Fig. 8.7a), there was no systematic change in these winter concentrations but the Esthwaite Water results (Fig. 8.7b) showed a rising trend. Figure 8.7c shows the result of using simple linear regressions to de-trend the two time-series. Once the increase associated with the changing trophic status of the lakes has been removed, the time-series are closely correlated ($r = 0.59$; $p < 0.001$) and the residuals from the two regressions can be used to quantify the effects of the observed variations in the weather. The same approach was used to de-trend the results acquired during the Q5 period, but here we used the average summer biomass of phytoplankton as the dependent variable.

8.3.3.3 The Influence of Winter Weather on the Supply of Phosphate

In 2004, George et al. published a paper that showed that the main factor influencing the winter concentration of phosphate in Blelham Tarn and Esthwaite Water was the inter-annual variation in the winter rainfall. Figure 8.8a and b show the relationship between the de-trended concentrations of phosphate in the two lakes during the winter (Q1) period and the recorded variation in the rainfall. The strongest correlation was that observed in Blelham Tarn ($r = 0.57$, $p < 0.001$) but a very similar pattern was recorded in Esthwaite Water ($r = 0.43$, $p < 0.02$). The most likely explanation for these correlations is the effect that heavy winter rains have on the composition of the surface drainage (George et al., 2004). When the winter rainfall is light, much of the dissolved phosphorus is adsorbed as the water passes through the upper layers of the soil (Sharpley and Sayers, 1979). In contrast, heavy rain increases the proportion of water that reaches the lakes as an overland flow which contains higher concentrations of dissolved phosphorus (McDiffet et al., 1989).

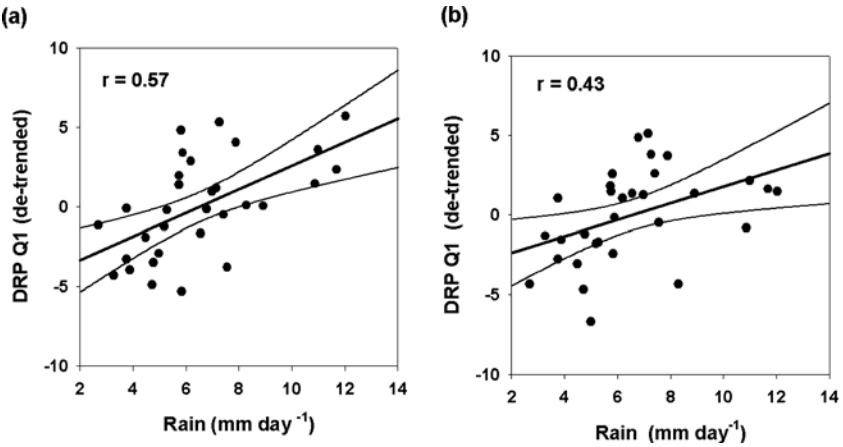


Fig. 8.8 The impact of year-to-year variations in the winter (Q1) rainfall on the de-trended concentrations of phosphate (μg P/l) in (**a**) Blelham Tarn and (**b**) Esthwaite Water

8.3.3.4 The Influence of the Summer Weather on the Recycling of Phosphorus

In these productive, thermally stratified lakes, the main factor influencing the internal recycling of nutrients is the wind-induced entrainment of water from the anoxic hypolimnion. In late summer, high concentrations of phosphate accumulate in the hypolimnion and can be entrained during-periods of intense mixing. These episodic transfers of phosphate from the hyplomnion into the epilimnion are difficult to detect in the raw time-series but become obvious when the data are de-trended to minimize the effects of enrichment and biological uptake. Figure 8.9 shows the relationship between the de-trended concentrations of phosphate recorded in the two lakes during the autumn (Q5) period and the average wind speed recorded during the same period. In Blelham Tarn (Fig. 8.9a), the highest phosphate concentrations were generally recorded in windy years, but the calculated correlation was not statistically significant. In Esthwaite Water (Fig. 8.9b), the relationship between the two variables was statistically significant ($r = 0.52$, $p < 0.001$), a response that can be explained by the size of the lake and its increased exposure to wind-induced mixing. Systematic differences of this kind show that lakes that are biologically similar but are very different in size may respond in different ways to changes in the weather. Thus small lakes, like Blelham Tarn and Esthwaite Water are likely to be more responsive to the short-term changes in the weather than large lakes that integrate the imposed climatic signal over longer time-scales. More detailed accounts of the climatic sensitivity of different types of lakes are given in Chapters 16 and 17 which also discuss the atmospheric features that regulate the regional and supra-regional coherence of lakes.

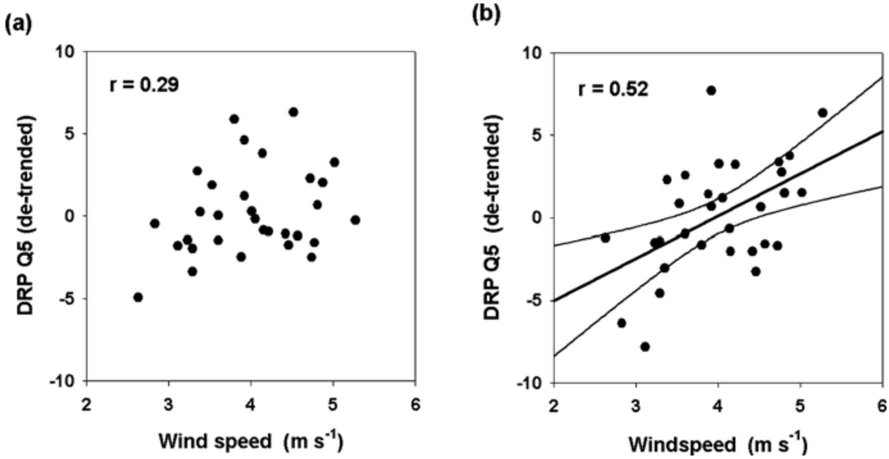


Fig. 8.9 The impact of year-to-year variations in the autumn (Q5) wind-speed on the de-trended concentrations of phosphate (μg P/l) in (**a**) Blelham Tarn and (**b**) Esthwaite Water

8.4 Discussion

The Case Studies assembled here demonstrate some of the ways in which year-to-year variations in the weather can influence the supply and internal recycling of phosphorus in lakes. As the world becomes warmer, these weather-related effects will become increasingly important and could have a profound effect on the growth and composition of the phytoplankton. The checklist in Table 8.2 summarizes the key weather-related effects observed in the four CLIME lakes, ranked according to their relative importance.

In winter, the most important effects were those associated with the change in the local air temperature and the rainfall. The air temperature effects were most pronounced in the northern lakes that freeze in the winter whilst the rainfall effects were more important in the lakes located in the milder western area. The warmer weather experienced during the last 10–15 years has caused a dramatic change in the hydrological flow pattern in northern Europe. The ice and snow cover on lakes in this region has been greatly reduced in both duration and thickness and there has been a corresponding advance in the timing of snow-melt. In Lake Erken, with its very long residence time, these flushing effects are less important and it is mainly the difference in the underwater light regime that has influenced the growth of phytoplankton and the associated uptake of phosphate. In shallow Lake Võrtsjärv, the residence time is much shorter, so the change in the nature and the timing of the precipitation has had a more pronounced effect on the water levels and the external loading. Once this lake is free of ice, the inter-annual variations in the intensity of wind mixing play an increasingly important part in the dynamics of this shallow lake. Short periods of intense mixing can then have a direct effect on the underwater light climate and nutrient concentrations and an indirect effect on the growth of phytoplankton. The western lakes were both relatively small and had very short residence times. Since they only freeze for a few days in the year, there were no

Table 8.2 The relative effect of the selected climatic variables on the supply and recycling of phosphorus in the four CLIME lakes

Climatic factor	Lake Erken	Lake Võrtsjärv	Esthwaite Water	Blelham Tarn
Winter air temperature	√√√	√√√	√	√
Winter rain/snow	√	√√	√√	√√√
Winter wind speed	√	√	√	√
Summer air temperature	√√√	√√√	√	√
Summer rain	√	√√	√√√	√√√
Summer wind speed	√√√	√√√	√√√	√√

ice and snow effects but the increase in the winter rainfall (see Chapter 19) had a significant effect on the supply of phosphorus which, in turn, influenced the growth and succession of the spring phytoplankton.

In summer, the observed variations in the air temperature and the wind speed had a significant effect on the internal recycling of phosphorus in all four systems. In Lake Erken, the associated increase in the stability of the lake influenced both the concentrations of phosphorus found in the hyplimnion and its entrainment into the epilimnion. In Lake Võrtsjärv, the critical mediating factor was the water level, so the summer supply of phosphorus was strongly influenced by the rainfall. The two western lakes responded in very similar ways to the observed variations in the summer weather but the mixing effect of the wind was only statistically significant at the larger, more exposed, site.

In this chapter, we have used a Case Study approach to explore some of the climatic factors influencing the supply and internal dynamics of phosphorus in four, very different, lakes. These processes are notoriously difficult to quantify, but these historical analyses have at least drawn our attention to the key physical drivers. A more detailed analysis of the way in which the projected changes in the climate might influence the external supply of phosphorus to these lakes is given in Chapter 9, this volume. Here, we have emphasized the role played by the internal re-cycling of phosphorus and the combined effects of seasonal variations in the air temperature, the rainfall and the wind. The processes that regulate the transfer of phosphorus from the sediment into the water column are influenced by a number of physical, chemical and biological factors (Blenckner, 2005). In thermally stratified lakes, the proportion of the inflowing P re-circulated in the open water is often quite small but these small fluxes have a disproportionate effect on both the growth and composition of the summer phytoplankton.

To date, there have been very few attempts to quantify the impact of short-term changes in the weather on the seasonal dynamics of phosphorus in either well-mixed or thermally stratified lakes. A variety of meteorologically driven processes are known to be involved and include the ‘pumping’ action of the internal seiche (Hutchinson, 1957) and the transfer of phosphorus from the littoral zone by convective and wind-induced water movements (George, 2000b). These re-cycling processes are, potentially, much easier to quantify in shallow lakes where there is more direct contact between the sediment and the open water. The example in Fig. 8.10 is taken from a recent study on Lake Võrtsjärv where the seasonal and inter-annual variations in the water level had a major effect on the annual phosphorus budget. In this study, the frequent monitoring of both the inflow and in-lake concentrations of phosphorus and nitrogen made it possible to quantify the effect of some extreme variations in the weather on the retention of nutrients in a rigorous way. The comparisons drawn between two years (1990 and 1995) showed very clearly how the retention of phosphorus was regulated by the combined effects of the winter rainfall and the water level fluctuations (Huttula and Nõges, 1998). Previous studies on this system have shown that the water level in Lake Võrtsjärv is positively correlated with the North Atlantic Oscillation (NAO) winter index (Järvet, 2004). In recent years, the NAO has tended to remain in its positive, ‘wet winter’ phase. If this trend

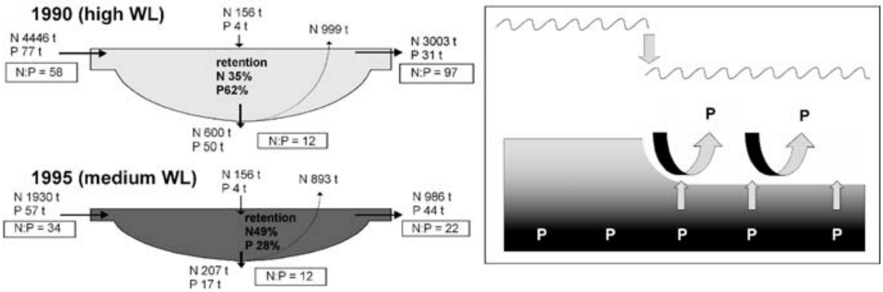


Fig. 8.10 The phosphorus budget produced for Lake Võrtsjärv to explain how the year-to-year variations in the water level influence the retention of P (redrawn from Nõges and Järvet, 1998 and Nõges and Nõges, 2004).

were to continue, then the internal loading of phosphorus to this lake would also be greatly reduced.

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Chapter 9
Modeling the Effects of Climate Change on the Supply of Phosphate-Phosphorus

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9.1 Introduction

The transfer of phosphorus from terrestrial to aquatic ecosystems is a key route through which climate can influence aquatic ecosystems. A number of climatic factors interact in complex ways to regulate the transfer of phosphorus and modulate its ecological effects on downstream lakes and reservoirs. Processes influencing both the amount and timing of phosphorus export from terrestrial watersheds must be quantified before we can assess the direct and indirect effects of the weather on the supply and recycling of phosphorus. Simulation of the export of phosphorus from the terrestrial environment is complicated by the fact that it is difficult to describe seasonal and inter-annual variations by existing process-based and empirical models. These variations are also strongly influenced by the history of the weather and by the frequency of extreme weather events. For example, the effects of storm runoff on the export of phosphorus can be very sensitive to levels of soil saturation and soil moisture, which are in turn influenced by the past history of precipitation and evapotranspiration. Inclusion of effects such as these is impossible using simple empirical models and difficult using process based models when model parameterization changes in response to antecedent conditions.

Using a purely empirical approach, phosphorus export is commonly estimated using fixed export coefficients based on local land use (i.e. kg P $km^{-2}y^{-1}$). These do not explicitly include any effects associated with local variations in the weather and cannot be used to explore climate related variations in the phosphorus supply. The models used to simulate the loss of phosphorus from watershed areas range from complex process based models to hybrid semi-empirical models. Schoumans and Silgram (2003) have reviewed some of the models used in Europe while Donigian and Huber (1991) have summarized some of the formulations used in the United States. In Europe, one of the most detailed process based models is the Systeme Hydrologique Europeen Transport model (SHETRAN Ewen et al., 2000). Other

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process-oriented models include the Soil Water Assessment Tool (SWAT Arnold et al., 1998), the Hydrologic Simulation Program Fortran (HSPF Bicknell et al., 1997), the Netherlands model (NL-CAT/AMINO Schoumans and Silgram, 2003), and the INCA-P model (Wade et al., 2002; Whitehead et al., 2007). These models rely on a combination of empirical, semi-empirical and process-based relationships to simulate some aspects of the phosphorus cycle, often with more detail on agricultural land. The advantage of these models is that they will simulate changes in the biogeochemical processes influencing phosphorus loss, and therefore should simulate climate related changes in both hydrologic and biogeochemical processes. These models often explicitly account for inputs of fertilizer and manure to agricultural lands, and maintain a soil phosphorus balance by simulating plant uptake, microbial decomposition, and phosphorus loss to surface and sub-surface flow. Greater levels of empiricism are found in other models of the phosphorus cycle i.e. REALTA (Schoumans and Silgram, 2003) or AGNPS (Young et al., 1986). Some models focus almost entirely on mechanistic descriptions of the hydrologic cycle, and use the observed relationships between concentration and flow to estimate the phosphorous losses associated with different hydrologic components. The HBV model is a widely used hydrologic model in Sweden and Europe (Bergström, 1995; Lindström et al., 1997) and has been used to estimate phosphorus export using regression equations linked to hydrologic flow components (Bilaletdin et al., 1994; Lidén, 2001; Arheimer and Lidén, 2000; Ulén et al., 2001). The Generalized Watershed Loading Function (GWLF) model used in the CLIME project is a good example of this hybrid approach. GWLF assigns fixed phosphorus concentrations to the surface runoff and base flow components of stream flow. A detailed description of this model has been given in Chapter 3 and further examples of its application are given in Chapters 11, 13 and 15.

A number of investigators have used climate models to drive hydrologic models that are then used to produce future estimates of watershed hydrological characteristics and changes in water yield (e.g. Lettenmaier et al., 1999; Nijssen et al., 2001), and see also Chapter 3. Studies which have used a similar approach to estimate the associated changes in phosphorus loading are not common, and most of these are based on the outputs from large-scale General Circulation Models (GCM). For example, Bouraoui et al. (2002) used a GCM coupled to the SWAT model to examine the effects of climate change on nutrient export from an agricultural watershed in England and found that the seasonality and magnitude of the changes varied greatly depending on the GCM used. Later, Bouraoui et al. (2004) used a similar approach to simulate long term total phosphorus loss from a Finnish watershed, and predicted that observed increases in phosphorus export could be attributed to decreased snow cover and increased winter runoff. More recently, investigators have coupled their watershed models to the higher resolution outputs produced by Regional Climate Models (RCMs). Arheimer et al. (2005) predicted the losses of nitrogen from a Swedish catchment using the HBV-N model driven by the same RCM as that used in CLIME (Rummukainen et al., 2004 and Chapter 2). Their results suggested an increase in N export ranging between 10 and 33%, while the hydrologic response between climate data sets varied greatly.

In this chapter, we describe the approach adopted by the CLIME project to quantify the effects of current climate and future climate change on the supply of phosphorus to lakes and reservoirs. In CLIME, GWLF was used to simulate the export of water and nutrients from a number of different watersheds located in Northern and Western Europe. GWLF was developed by Haith and Tubbs (1981) and validated by Haith and Shoemaker (1987) to simulate monthly dissolved, particulate and total phosphorus and nitrogen loads in stream flow. There are several adaptations of the original GWLF model currently in use (Dai et al., 2000; Lee et al., 2000; Schneiderman et al., 2002; Limbrunner et al., 2005; Mörth et al., 2007). This chapter explains how the model was used to estimate the flux of phosphate phosphorus and describes the methods used to quantify the uncertainties associated with these projections. Since the outputs from GWLF were used to drive models that required inputs of soluble phosphate phosphorus (Chapter 15), these are the results described here; however, the same model can also be used to simulate the transport of particulate phosphorus (Haith et al., 1992).

GWLF is a practical compromise between simple models based on empirical export coefficients (Reckhow, 1979; Reckhow and Chapra, 1983) and more complex models that simulate chemical processes. GWLF dynamically simulates variations in stream discharge and combines these results with source-specific concentrations of phosphate phosphorus to estimate phosphorus export This approach proved to be a good compromise between static export coefficients and complex descriptions of the phosphorus cycle. The model effectively captures the hydrologic processes responsible for regulating the export of phosphorus while maintaining the simplicity required to apply the model to a range of lakes situated in very different climatic regions.

A key objective in CLIME was to use a consistent watershed modeling approach to estimate nutrient export from a number of different watersheds covering a wide range of climate and physiography. This required the use of a model that was simple, flexible and could be adapted to meet the needs of the modelers in different regions within the 3 years allocated to the project. The version of GWLF used in CLIME was created by the New York City Department of Environmental Protection (Schneiderman et al., 2002) using the Vensim visual modeling software package (Ventana Systems, Inc. http://www.vensim.com). This version was particularly well suited for use in CLIME, since the graphical interface allowed the structure and governing equations of the model to be understood by all CLIME participants. This interface was easy to use and allowed the users to modify the model and gain a clear understanding of the how the model worked i.e. how inputs were transformed to outputs; how the model equations and parameters facilitated this transformation; and the sensitivity of model to variations in the key parameters. All these issues were explored in CLIME modeling workshops where we collectively modified and tested the model, developed common methods of calibration and common simulation strategies. As a consequence, CLIME was able to produced a set of consistent simulations and provide the research groups in the Member States with new skills in watershed modeling. This modeling competence can now be applied to other European applications and the model can be adapted to simulate other fluxes and responses to differing pressures.

9.2 The GWLF Soluble Phosphorus Model

The GWLF hydrologic sub-model (Chapter 3) is driven by daily temperature and precipitation data and water balances are calculated on a daily interval. Stream discharge is also output on a daily interval, and consists of direct runoff and base flow components. In GWLF the ‘loading function’ approach links fixed phosphorus concentrations to direct runoff and base flow as illustrated in Fig. 9.1. Phosphorous export in direct runoff is calculated by multiplying land-use specific direct runoff volumes by land-use specific dissolved phosphorous concentrations. Agricultural practices such as manure spreading can be accounted for by seasonally adjusting the concentration of dissolved phosphorus in direct runoff from agricultural lands. Base flow export is calculated as the product of base flow and a single dissolved phosphorus concentration assumed to be representative of base flow from the basin as a whole.

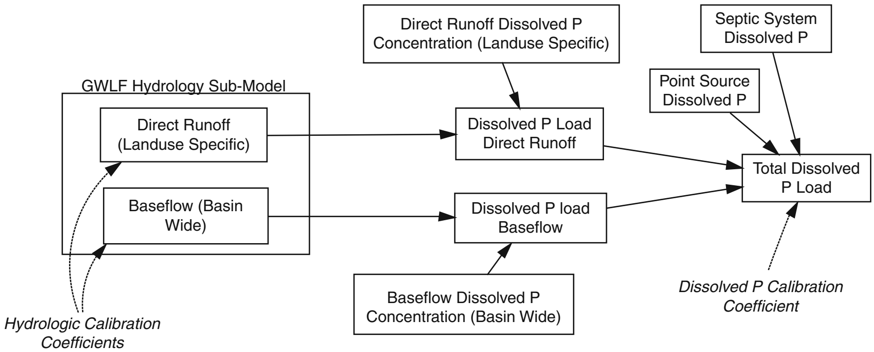

Fig. 9.1 Conceptual diagram illustrating the coupling of GWLF simulated hydrology to dissolved phosphorus export. Phosphorus export by direct runoff is land-use specific, while export occurring in base flow is estimated using a single basin-wide calculation. Septic and point source loads are also considered

The contribution of phosphorus from septic systems can also be estimated, based on population density and estimates of septic system performance. Point source loadings (e.g. wastewater treatment plants), can when present, be input in a variety of ways depending on the resolution of the point source loading estimates. Both septic and point source inputs can be added to the estimated total phosphorus load leaving the watershed.

In GWLF, the simulated changes in phosphorus export are primarily a function of the changes in watershed hydrology, which in turn regulate the total volume of water available to transport the phosphorus defined by the loading function concentrations. The model can still represent some of the processes that influence the concentration of phosphorus in the exported water. These processes can be divided into two classes: (1) hydrologic processes which influence the relative contribution of the different sources of phosphorus to the stream, and therefore the flow-weighted mean concentration in exported water and; (2) biogeochemical processes that influence the concentration of phosphorus in the waters associated with each source. GWLF can simulate the first set of effects but does not simulate the biogeochemical processes that modulate the source specific concentrations.

9.3 Overview of Regions

In CLIME, the sites used for phosphorus modeling were all located in Northern and Western Europe. Table 9.1 summarizes the physical characteristics of these sites whilst the map in Chapter 1 shows their geographic location. The two sites selected from Northern Europe (Arbogaån and Mustajoki) were very different in size, but were exposed to similar levels of precipitation, and had similar levels of stream flow relative to precipitation: The long-term mean ratio of stream flow/precipitation is 0.47 for the Arbogaån watershed and 0.40 for the Mustajoki watershed. In both of these watersheds the accumulation of snow and snowmelt play an important role in regulating the seasonal pattern of stream flow. Winter precipitation occurring as snow accumulates in the watershed until the spring, when snowmelt augments stream flow to form a distinct snowmelt peak that often results in maximum annual stream flow. The Finnish site has a more continental climate than the Swedish site since it is further from the Baltic Sea. Under current conditions, 29% of the annual precipitation at the Finnish site falls as snow whilst the corresponding value for Arbogaån is 23%. Contemporary Finnish air temperatures are also colder, with mean December−February temperatures being −6.9 C as compared to −2.6 C in the Swedish watershed.

Table 9.1 The basic characteristics of the CLIME sites discussed in this chapter

Country	Watershed	Position	Area (km ^{2})	Mean annual precipitation (cm)	Mean annual stream flow (cm)	Mean annual phosphate P export (kg km^{-2}y^{-1})
Northern region						
Finland	Mustajoki	61.04 N 25.08 E	76.8	72	29	2.91
Sweden	Arbogaån	59.59°N 16.6°E	3808	90	42	6.43
Western region						
Ireland	Flesk	52.03°N 9.33°W	332	210	152	13.76
U.K.	Esthwaite	54.60°N 3.10°W	15.6	221	176	11.32

The Western sites have a maritime climate with high levels of precipitation, very little snow and less seasonal variation in air temperature. The two regions have similar summer precipitation; however, at the Northern sites these correspond to the annual maximum whilst at the Western sites they are close to the annual minimum. At present, the air temperatures recorded during the winter, spring and autumn are all higher at the Western sites. The air temperatures at the Northern sites are, however, slightly higher during the summer. This leads to somewhat greater levels of evapotranspiration at the Western sites, but not enough to compensate for regional differences in precipitation. Since greater precipitation leads to higher levels of soil water storage and antecedent wetness, stream flow/precipitation ratios are much higher at the Western sites. Long term averages of this ratio range from 0.72 in Ireland to 0.79 in England.

The maximum levels of stream discharge consequently occur at the Western sites during the autumn and winter, and average stream flow levels are much higher being more than three times those recorded in the Northern region. These regional differences in precipitation and stream flow have a profound effect on both the magnitude and the seasonality of phosphorus export. Indeed under present day conditions, the simulated mean annual rates of phosphorus export in the Western sites was more than double that estimated for the Northern watersheds.

9.4 Calibration of GWLF Phosphorus Model

Using the GWLF model to estimate phosphorus loading requires calibration of the hydrology sub-model parameters as described in Chapter 3. After the hydrology calibration, a number of additional model parameters must be estimated, the most important of which are the phosphorus concentration in the direct runoff from the different land use categories and a basin-wide phosphorous concentration for base flow. If appropriate, additional information on potential sources such as watershed population, septic system performance, and estimates of point source inputs may also be used to parameterize the model (Fig. 9.1). Since site specific concentration data are rarely available, in most cases the fixed concentrations are derived from databases covering large (often national) geographic regions. These values (Table 9.2) were provided by the CLIME partners working on each watershed since they were more familiar with the type and quality of information available.

Given the inherent difficulties in estimating site specific concentrations from a disparate set of data, a final global calibration of the phosphorus model was performed. A single time invariant factor (Table 9.2) was used to simultaneously adjust the phosphorous concentrations associated with both direct runoff and base flow. This factor was adjusted so that the simulated long term mean phosphorus concentration in total stream flow matched the long term mean calculated from stream water measurements. The result was to constrain the variation in simulated phosphorus concentration to that measured under present conditions, while allowing

Table 9.2 The area of CORINE land use and the phosphate phosphorus concentration associated with direct runoff from each land use. At the bottom of the table the single groundwater concentration associated with each site is also shown. The concentrations shown in the table are calibrated. i.e. those obtained from national data sources and multiplied by the global calibration factor, also shown at the bottom of the table. Estimates of the proportion of the phosphorus load associated with different sources are based on a control period simulation

Land use classification CORINE class	Area (ha)				Phosphate phosphorus (mg l^{-1})			
	Mus FI.	Arb SE.	Flesk IE.	Esth UK	Mus FI.	Arb SE.	Flesk IE.	Esth UK
Continuous urban		58	1			0.256	0.043	
Discontinuous urban		3,188	188			0.179	0.043	
Industrial		958				0.256		
Road rail networks	10	168	170		0.002	0.256	0.043	
Airports		84				0.256		
Mineral extraction		348				0.179		
Dump sites		200				0.179		
Construction sites		12	380			0.256	1.207	
Green urban areas		2,036				0.023		
Sport leisure facility		814	65			0.179	0.013	
Non irrigated land		44,576				0.135		
High prod pasture			2,681	867			0.019	0.012
Mixed prod pasture		6,835	2,286			0.135	0.013	
Low prod pasture			5,170				0.013	
Complex cultivation	998		99		0.023		0.013	
Mainly agriculture			842				0.013	
Broad leaved forest		13,446	135	210		0.006	0.008	0.004
Coniferous forest	5,146	174,426	1,658	315	0.005	0.006	0.008	0.004
Mixed forest		25,255				0.006		
Natural grasslands			3,064	170			0.008	0.003
Moors heathland			5,622				0.008	
Transition woodland		66,457	57			0.006	0.008	
Sparsely vegetated			348				0.008	
Bare rocks		2				0.004		
Inland marshes		821	92			0.004	0.008	
Exploited peat bogs			238				0.008	
Unexploited peat	1,536	1,4053	9,092		0.013	0.004	0.008	
Water courses		649				0.004		
Water bodies		26,389	346	99		0.004	0.008	0.000

	Mus FI.	Arb SE.	Flesk IE.	Esth UK
Ground water Concentration (mg l^{-1})	0.010	0.007	0.006	0.007
Global calibration factor	1.56	0.51	1.27	0.45
Percent load direct runoff	32.5	17.9	49.0	40.4
Percent load base flow	67.5	69.2	27.7	59.6
Percent load septic systems[1]	0	0	19.8	0
Percent load point source	0	12.9[1]	3.5	0

[1] In Sweden septic loads were available as daily areal loading estimates. These were added to the point source loads.

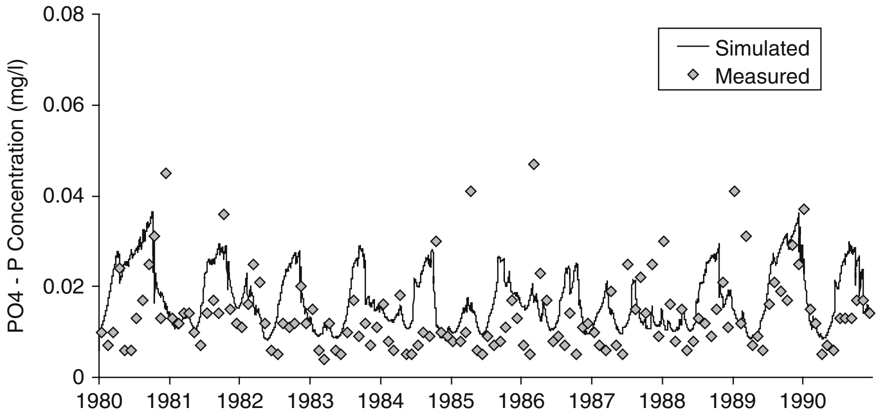

Fig. 9.2 Simulated and measured dissolved phosphorus concentration for the Arbogaån watershed in Sweden following global optimization of the GWLF phosphorus model

temporal variations to occur as a consequence of the projected changes in runoff and base flow.

The most pronounced variations in phosphorous concentration were those recorded in the large Swedish watershed where there was a greater variety of land use types and land use-specific phosphorus concentrations. (Fig. 9.2). The simulated variations in phosphorus concentration in the other catchments were less pronounced and were less than the variability in measured concentrations. In all cases however, the main goal of the optimization was met i.e. the simulated variation in the phosphorus concentrations were adjusted to realistic levels and were held within the range of variation associated with historical measurements.

Comparison of the optimized phosphorus concentrations (Table 9.2) shows good agreement between sites for the major land use categories and base flow concentrations, despite the fact that concentrations were assigned from different regional databases. There were large differences in the site-specific concentrations associated with urban and suburban lands, however, these areas were small and had little effect on the export of phosphorus. Partitioning the phosphorus load between direct runoff and base flow showed regional differences with a greater proportion of the load appearing in surface runoff in the much wetter Western Region. In the Irish watershed, the input from septic tanks was proportionately greater (Table 9.2), and errors in estimating these inputs at times of low flow can lead to overestimation in summer concentrations (Whitehead et al., 2007).

Since GWLF does not dynamically simulate variations in phosphorus concentration, the criteria used to judge the success of the optimization can be less rigorous than those used for the hydrology sub-model decribed in Chapter 3. The data in Fig. 9.2 show that the simulated variations in phosphorus concentration broadly reflect the seasonal variation in the inputs from different sources. However, it would not be realistic to expect these simulated concentrations to closely

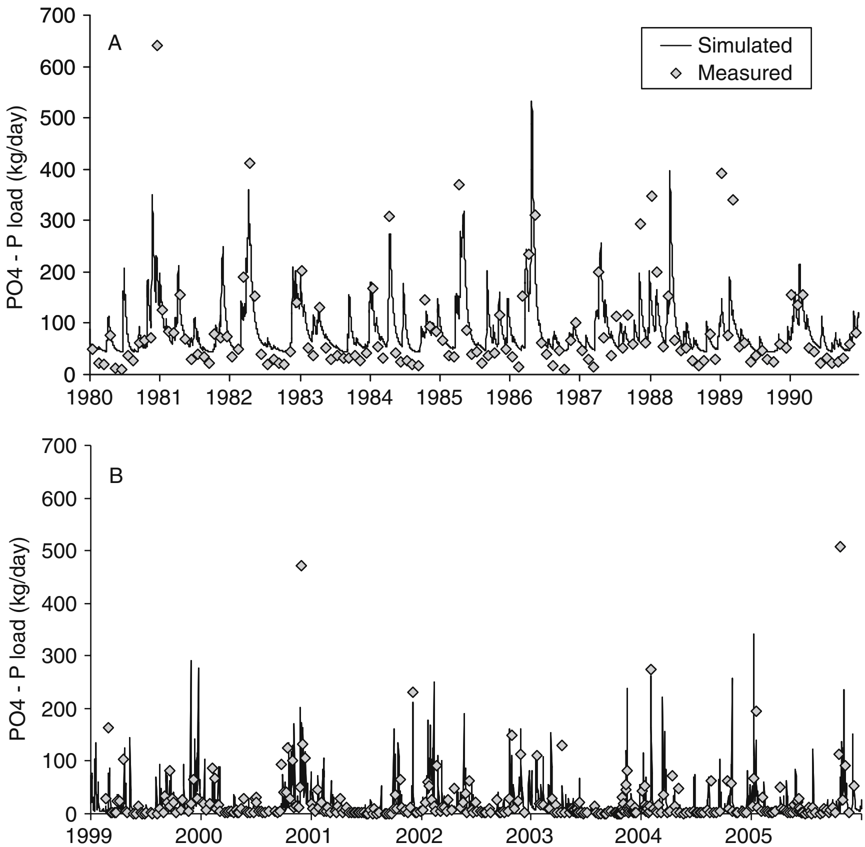

Fig. 9.3 Simulated dissolved phosphorus loads compared with measured daily loads for Arbogaån in Sweden (**A**) and Flesk in Ireland (**B**). Measured loads are calculated as the product of a stream water phosphorus measurement and mean daily discharge on the day of measurement

correspond to actual variations in measured stream water concentration, since measured data would be influenced by a variety of additional processes.

In Fig. 9.3 simulated daily phosphorus loads are compared to the instantaneous loads calculated as the product of the measured phosphorus concentration and the discharge on the day of measurement. These data show a better correspondence between simulated and measured loads since variations in simulated hydrology plays a dominant role in regulating the simulated variations. Judging the fit of such phosphorus loading simulations by statistics such as the Nash Sutcliffe statistic (Nash and Sutcliffe, 1970), is difficult unless sampling during the calibration/verification period is frequent enough to provide reliable estimates of phosphorus loading i.e. including storm events. The Irish site did have frequent and adequate sampling for estimating phosphorus loads, and at this site, the Nash Sutcliffe statistic for the 1999–2004 period was 0.57.

9.5 Simulations and Uncertainty

Simulations of the future climate have inherent uncertainty over a range of temporal and spatial scales. Part of this uncertainty is associated with the climate model used and part with the methods used to downscale these results to the area of interest (Chapter 2, this volume). Much of the uncertainty in predicting future phosphorus loads is thus directly linked to the uncertainty in the input climate data. To assess this uncertainty, simulations were made using multiple climate models, two emission scenarios, and two techniques for statistical downscaling. Table 9.3 lists the models and emission scenarios used to produce multiple realizations of the weather at a given site. The two Regional Climate Models used (RCAO and HadRM3p) were driven by two Global Climate Models (HadAM3 and ECHAM4) and perturbed by two levels of greenhouse gas emission (IPCC A2 and B2) The control simulations were for the 1961–1990 reference period recommended by the IPCC and the future simulations covered the 2071–2100 period. At the Western sites the RCM outputs were downscaled using the weather generator described in Chapter 2. At the Northern Sites in Sweden and Finland, where this approach was not considered appropriate the outputs were downscaled using the delta change method explained in Chapters 2 and 13. By providing multiple (100–500) '30 year' realizations of the control and future simulations, these techniques allowed us to estimate the uncertainty derived from expected variations in weather and also quantify the potential effects of the more extreme variations that might occur.

Table 9.3 (A) Combination of climate models and emission scenarios that were used to drive GWLF simulations. For each GCM/RCM combination a control (present climate) scenario was also run. (B) Method used to produce multiple weather scenarios for the CLIME study sites and the number of realizations produced at each site

(A) Summary of climate simulations

Abbreviation	GCM	RCM	IPCC emission scenario
HA2	HadAM3H/P	RCAO	A2
Had A2	HadAM3H/P	HadRM3p	A2
EA2	ECHAM4/OPYC3	RCAO	A2
HB2	HadAM3H/P	RCAO	B2
Had B2	HadAM3H/P	HadRM3p	B2
EB2	ECHAM4/OPYC3	RCAO	B2

(B) Methods of generating multiple local weather sequences

CLIME site	Method used to simulate local weather	Number of realizations
UK-Esthwaite	Weather generator	500
Ireland-Flesk	Weather generator	100
Sweden-Arbogaån	Re-sampling	100
Finland-Mustajoki	Re-sampling	100

9.6 Changes in Phosphorus Export over Annual Time Scales

Figure 9.4 summarizes the differences in phosphorus loading simulated for the CLIME sites in the two regions where the loads are estimated on an annual basis and normalized by watershed area. There are large differences between the Northern

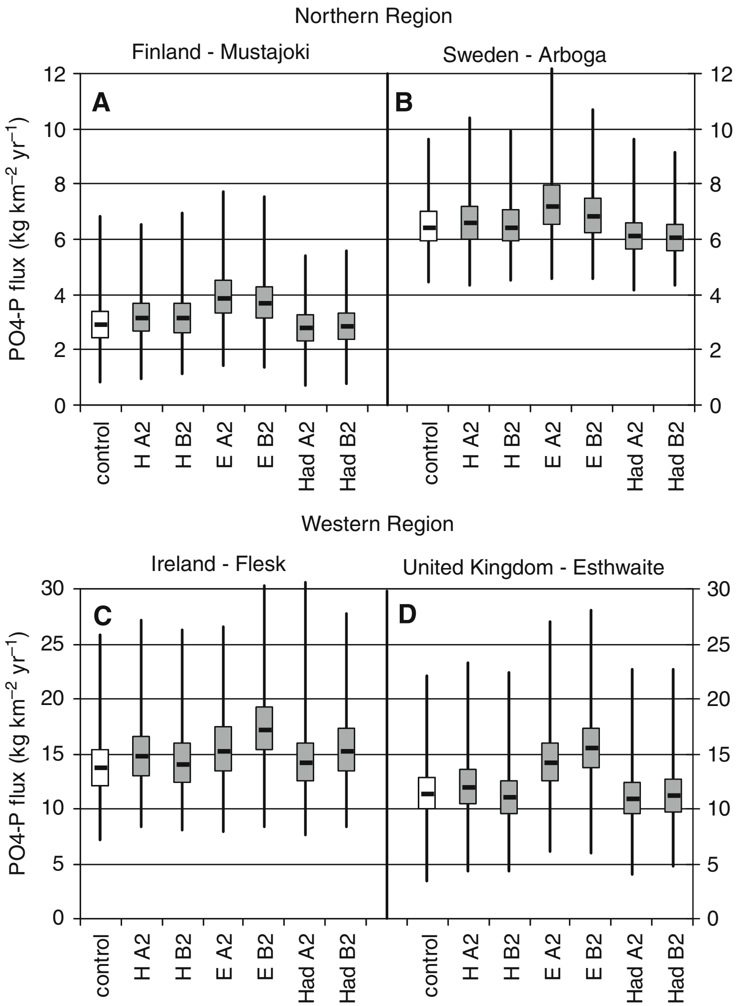

Fig. 9.4 Mean annual rates of phosphorus export calculated for (**A**) Finland, (**B**) Sweden, (**C**) Ireland and (**D**) United Kingdom study sites. These plots summarize the results from GCM/RCM simulations perturbed by the A2 and B2 emission scenarios. The plots show the range of variability between the 5 and 95 percentiles, the 25 and 75 percentiles, and the median values where the variability is that resulting for the multiple realizations of the daily weather

and Western Regions, which are largely attributed to the regional differences in precipitation and stream flow summarized in Table 9.1, and described in detail in Chapter 3. The larger phosphorus loading rates simulated for the Western sites are the result of the much higher levels of precipitation and the increased ratio of stream flow to precipitation at these sites. Intra-regional variability in phosphorus export is related to these differences in hydrology, and also to differences in watershed physiography, land use and biogeochemistry.

The results in Fig. 9.4 and Table 9.4 show that the projected changes in the climate led to very little overall change in the annual loading of phosphorus at the selected sites. In most cases there was a small to moderate increase in annual phosphorus loading with the most pronounced increases being recorded with the Max Planck ECHAM model. Simulations with the Hadley Centre model typically produced small reductions in the phosphorus load but moderate increases were recorded in the Flesk watershed in the west of Ireland (Table 9.4).

In both regions seasonal changes in stream flow (Chapter 3) resulted in increased discharge and phosphorus loss in the winter and fall, which was roughly balanced by a decreased loss in the summer and spring. In general the projected differences between the present and future simulations were of the same magnitude as the variability associated with the choice of RCM and the variance generated by the multiple realizations of the daily weather.

Table 9.4 Annual percent change in median daily dissolved phosphate phosphorus loads for the four CLIME sites

	Climate model/scenario					
Site	H A2	H B2	E A2	E B2	Had A2	Had B2
Mustajoki, Finland	+9	+7	+33	+27	−5	*−2*
Arbogaån, Sweden	+2	0	+11	+6	−5	*−6*
Flesk, Ireland	+7	+2	+11	+25	+3	*+11*
Esthwaite, UK	+6	−3	+25	+37	−4	*−1*

9.7 Seasonal Patterns in Phosphorus Export

Figure 9.5 shows the observed and projected seasonal variation in the export of phosphorus from the four CLIME sites. The solid line shows the pattern based on the median of the simulations for the control period and the shaded areas the range of values simulated for the future scenarios listed in Table 9.2. A comparison of the results from the two regions shows two contrasting patterns of seasonal variation.

In the Northern Region snow accumulation and snowmelt have an important influence on the timing of phosphorus export, as can be seen from the simulated current patterns of export in Sweden and Finland. Cold temperatures and snow accumulation limits winter phosphorus export, whilst snowmelt augments the spring runoff

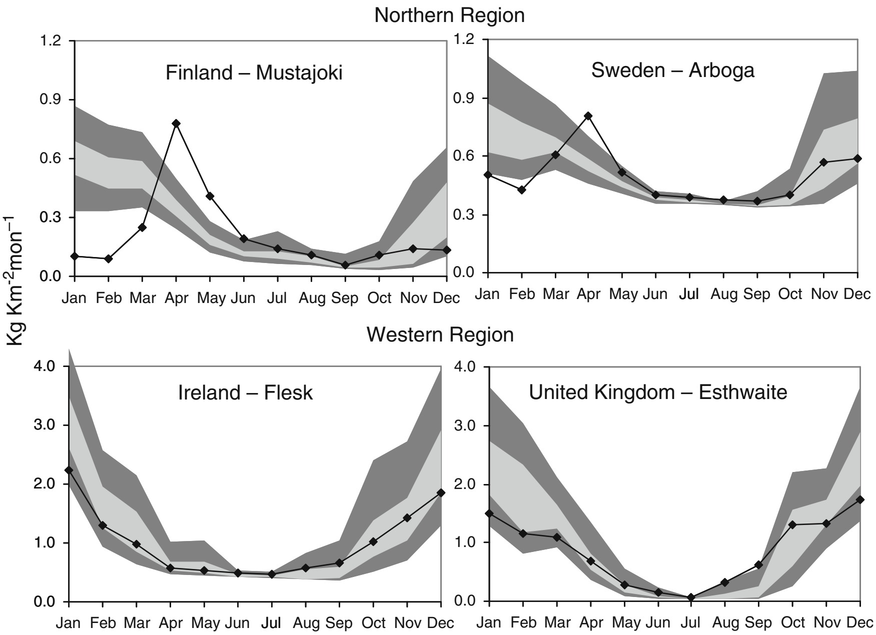

Fig. 9.5 Simulated seasonal patterns in phosphate phosphorus export based on all the simulations listed in Table 9.3. The line shows the median of the control (present day) simulations, while the *shaded area* shows the range of variation associated with all future scenarios. The *light gray* shading shows the range of the medians and the *dark gray* shading the range between the 25 percentile and 75 percentile limits in the six future climate simulations

and leads to maximum export rates. Under future conditions, the rate of warming is projected to be most pronounced in these northern areas, with the largest increases projected for the winter months. This results in greater amounts of rain and lower amounts of snow during the winter, with concomitant decreases in snow accumulation and increased winter stream flow. A number of other studies have predicted similar changes in the seasonality of stream flow for Scandinavian (Vehviläinen and Lohvansuu, 1991; Bergström et al., 2001; Andréasson et al. 2004; Graham, 2004; Kaste et al., 2004) and other Northern watersheds (Rango, 1995; Nijssen et al., 2001; Barnett et al., 2005). All future climate scenarios suggest that there will be a profound change in the export patterns recorded during the winter to spring period: Winter export will be very much greater while the spring peak in phosphorus export will gradually disappear. This is by far the most important change to the in phosphorus loadings projected for the CLIME sites. We can also have a high degree of confidence in the projected changes since the 25–75 percentile range of future phosphorus loads fall outside the present day values for the winter to spring period (Fig. 9.5). The most pronounced change in the timing of phosphorus export is projected for the site in Finland, since the amount of snow that currently accumulates here is greater and more persistent than at the site in Sweden. Arheimer et al. (2005) and

Kaste et al. (2004) have described very similar seasonal shifts in the pattern of nitrogen transport that were related to changes in stream flow for watersheds in Sweden, Norway, and Finland, where snow accumulations and snowmelt play an important role in the present day hydrological cycle.

In the Western Region both present and future simulations show a similar seasonality in rates of phosphorus export, with maximum loads occurring in the winter months, followed by a steady decline through the spring to minimum values in mid summer. These trends are also related to the projected seasonal variations in stream flow. Future scenarios predict that increases in stream flow will occur in the winter and fall (Chapter 3), further accentuating the trends already identified in our historical analyses (Chapters 18 and 19). Comparing the phosphorus export rates for control and future scenarios in Fig. 9.5, suggests that in most future years there will be greater rates of phosphorus export in the winter period, since the range of future loads generally lies above the current rates of export.

9.8 Seasonal Changes in Phosphorus Export

Figure 9.6 shows the differences between dissolved phosphate phosphorus loadings simulated by the control runs and the future runs of the climate models. In this figure, the calculated differences are summarized by region, and the effects of the A2 and B2 emission scenarios are examined separately. The projected changes for the Northern and Western Regions show that, even though there is little change in the annual rates of phosphorus export (Fig. 9.4), there are important changes in the timing of the loadings which can be directly related to the projected changes in stream flow.

In the Northern Region the simulated future changes in phosphorus loading were almost entirely associated with changes in winter precipitation. More frequent winter snowmelt, and an increase in the proportion of winter precipitation that fell as rain, led to increased winter stream flow and phosphorus export which led to a significant shift in the seasonal timing of nutrient delivery. The peak loads shifted from occurring in correspondence with spring snow melt, to more widespread occurrences spread out over the entire winter (Figs. 9.5 and 9.6). Given the greater importance of snow accumulation and snowmelt in the Finnish catchment, the difference graphs in Fig. 9.6 show the changes described as being more pronounced for Finland than Sweden.

In the Western Region there is little difference in the rates of phosphorus export in the early part of the growing season (May–July), but there are significant changes in the loadings simulated at other times of the year (Fig. 9.6). From late summer and through the fall there was a general decrease in the rate of phosphorus export at both Western Sites, but this decrease was less pronounced and of a shorter duration at the site in Ireland. In winter, there was an increased loading at the UK site and for the Irish simulations driven by the ECHAM model. The Irish simulations driven by

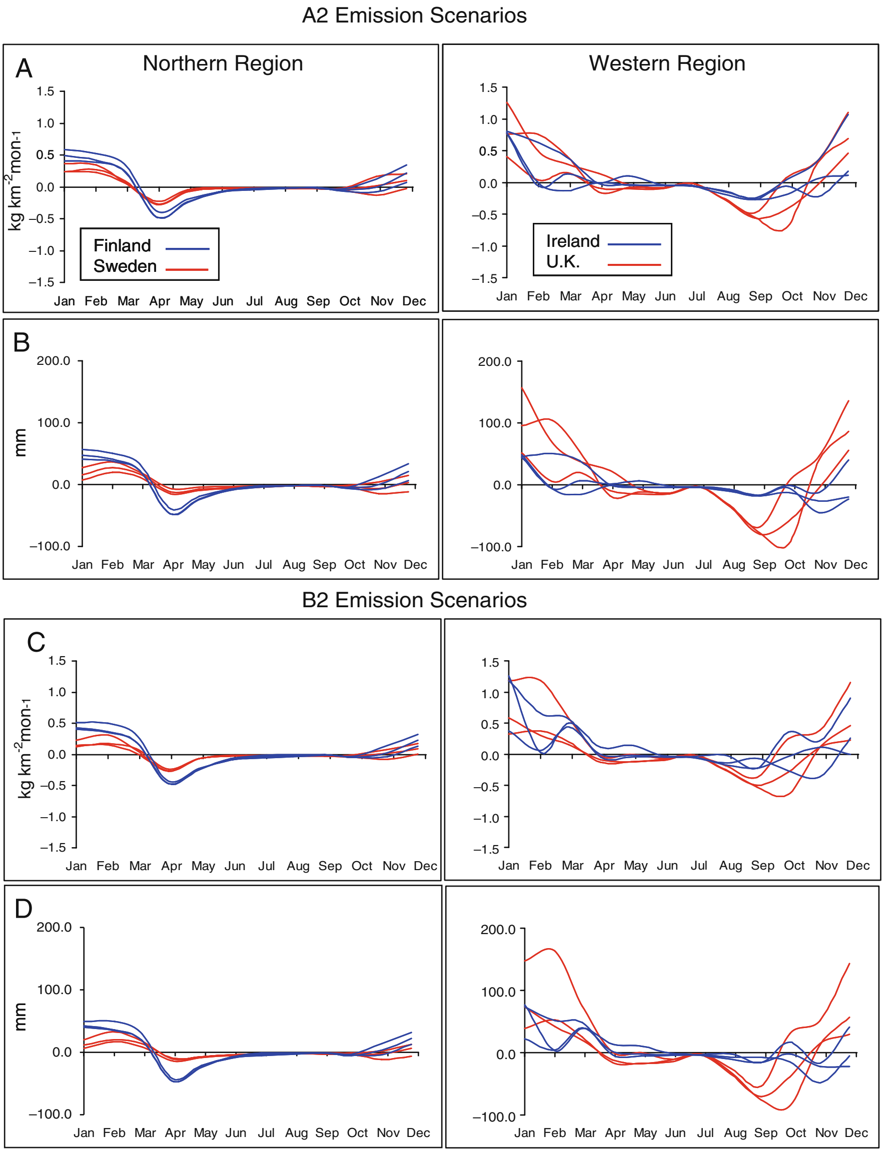

Fig. 9.6 Difference between the future and control simulations for the three GCM/RCM model runs. Difference values are calculated from the median values of mean monthly rates of phosphorus export, based on multiple simulations for each GCM/RCM/emission scenario. Differences in phosphorus export are shown in A and C for the A2 and B2 emission scenarios. Corresponding differences in stream discharge are shown in B and D

the Hadley Centre model produced a different result with a modest decrease in the winter loading. Similar patterns of change in UK stream flow have been reported (Arnell, 2003; Arnell and Reynard, 1996) and these patterns of change are strongly related to simulated increases in precipitation between October and February, and simulated increases in evaporation which were greatest between May and September. Such changes in precipitation and evapotranspiration influence soil water capacity, soil water recharge, and the levels of stream flow and nutrient export during the summer and fall months as already noted in Chapter 3.

In general, the differences associated with the two emission scenarios were less than those observed for the two climate models. The B2 scenarios in Fig. 9.6 show much the same patterns as those discussed above for the A2 scenarios but the projected changes were slightly less pronounced. These results underline a feature already noted in Chapter 2. Here, Samuelsson quotes the results of a study by Déqué et al. (2007) which showed that the greatest source of uncertainty in the current generation of climate models was the choice of GCM. The choice of RCM had less of an effect on the overall variability but had as large an effect on summer precipitation as the choice of GCM.

9.9 Conclusions

There are numerous investigations of the direct effects of climate change on lakes, particularly studies of thermal effects related to water temperature and stratification (e.g. De Stasio et al., 1996; Fang and Stefan, 1999; Blenckner et al., 2002), and effects on the timing and duration of ice cover (e.g. Magnuson et al., 2000; Yoo and D'Odorico, 2002 and Chapter 4, this volume). There are also some studies which consider the effects of changing climate on the trophic structure of lakes (e.g. Weyhenmeyer, 2001; Malmaeus et al., 2006; Markensten et al., 2009 and Chapter 14, this volume). However, there are few studies which simultaneously examine the effect of climate change on watershed hydrologic processes regulating nutrient delivery and lake processes regulating thermal structure, mixing and temperature dependent biogeochemical processes.

Coupled watershed and lake models have been used to estimate the effects watershed management and in particular the effects of agricultural best management practices on downstream water bodies (e.g. Mankin et al., 1999; Mankin et al., 2003; Summer et al., 1990). Owens et al. (1998) describe a system of linked watershed – reservoir models that can be used to evaluate changes in watershed management on the water quality of reservoirs of the New York City water supply. Wulff et al. (2007) has used a coupled modeling system, including a variation of the GWLF watershed model, to evaluate effects of changes in agricultural practices land use and phosphorus management on Baltic Sea trophic status.

Studies using coupled watershed and lake models to evaluate the effects of climate change are not as common. In Sweden both Arheimer et al. (2005) and

Markensten et al. (2009) have used coupled watershed lake models to evaluate the effects of projected future changes in the timing and magnitude of nutrient loading on lake biomass and trophic structure. A recent study by Komatsu et al. (2007) in Japan coupled a simple watershed water quality model to a lake model that simulated future changes in sediment geochemistry and phytoplankton biomass. Whitehead et al. (2009) also coupled watershed water quality simulations to simulations of aquatic biology; although in this case the effects of climate change on the marcophyte and epiphyte vegetation in large river systems were examined.

An important strength of the phosphorus scenarios presented here is that these simulations were driven by climate scenarios derived from combinations of different global and regional climate models, which were in turn driven by two different emission scenarios (Table 9.3). This provided an ensemble of phosphorus loads for the control period and future scenarios, which allowed both the magnitude and variability of future changes to be examined on both annual and seasonal time scales. Many of the simulations presented in this chapter suggest that, on an annual basis, little change in the magnitude of phosphate phosphorus loading will occur as a result of climate change (Table 9.4). However, it is significant that nearly all simulations suggest an increase in annual phosphorus loading will occur, and that a few of the simulations suggested that this increase could be as large as 30%.

More importantly, the simulations presented here suggest that even if there are only small changes in the annual rates of phosphorus loading, changes in the seasonal timing of phosphorus delivery to lakes will occur, and that the certainty in these changes can be predicted with far less ambiguity (Fig. 9.5). In both Northern and Western regions increased loads are projected during fall and winter. At the Western sites contemporary seasonal patterns of phosphorus loading will be accentuated and result in increased phosphorous loads between October and March. At the Northern sites a profound shift in the timing of loading is projected to occur, with greater rates of loading in the winter occurring at the expense of a greatly reduced load during a time that was previously associated with spring snowmelt. For both Western and Northern sites this will result in a greater proportion of phosphorus being transferred to lakes during conditions of low light, isothermal mixing, and cold temperatures, all of which do not favor phytoplankton growth. GWLF does not simulate changes in biogeochemistry that would affect source specific phosphorus concentration. However, predicted future shifts in the seasonality of the phosphorus loading suggest that watershed biogeochemical processes could also be of importance, since greater phosphorus loading would occur during the winter when uptake by terrestrial vegetation would be less.

The projected reduction in phosphorus export during the spring at the Northern sites could lead to significant changes in the growth and seasonal succession of phytoplankton. Spring is a time in which the input of phosphorus has the most direct effect on the growth of phytoplankton, especially if the timing of spring runoff is considered in relation to other climate mediated processes influencing the loss of lake ice cover, and the onset of thermal stratification. Changes in the timing of phosphorus loading during the spring could result in a marked shift in the timing of the spring bloom even if the maximum biomass attained was constrained by the reduced loading.

The major conclusion of this chapter is that changes in the timing of phosphorous export rather than changes in the magnitude of export are the most significant expected consequence of future climate change on lakes in Europe. These shifts in the timing of the phosphorus fluxes can lead to important changes in lake trophic status and phytoplankton succession. A major strength of the CLIME project was the way in which the same system of coupled models was used to systematically explore these effects over a range of very different lakes. The multiple realizations of the seasonal cycles summarized here also allowed us to quantify at least some of the uncertainties associated with the different climate models and the contrasting emission scenarios. The methods used to assess the impact of climate change on natural systems are currently undergoing a shift away from simple deterministic techniques towards more realistic stochastic procedures. The GWLF model used here proved particularly effective at generating the very large number of outputs required to quantify the uncertainties associated with the rapidly changing climate.

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Chapter 10
The Impact of the Changing Climate on the Supply and Recycling of Nitrate

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10.1 Introduction

A high proportion of the nitrogen found in lakes and rivers is present in the form of nitrate. The concentration of nitrate in many surface waters has increased over the last forty years (OECD, 1982; Roberts and Marsh, 1987; Johnes and Burt, 1993). The main source of nitrate is diffuse drainage from agricultural land (Vinten and Smith, 1993) but point sources can be important in populated areas (Jarvey et al., 1998). In 1991, the European Union introduced the Nitrates Directive (91.676) to protect waters from pollution by nitrate leached from agricultural land. Over the years, a number of scientific projects have been funded to provide strategic support for this directive. These include the Euroharp project (www.euroharp.org) that compared budgeting methods and the INCA project (www.rdg.ac.uk/INCA) that developed a new simulation model.

A number of approaches have been used to quantify the factors influencing the transfer of nitrate from terrestrial to aquatic systems. These include budget calculations (Reynolds et al., 1992; Johnes, 1996), process-based models (Wade et al., 2002; Tipping et al., 2006) and statistical analyses (Hirsch et al., 1982; Easterby, 1997). Most budget based studies rely on measurements acquired over short periods of time (Johnes and Burt, 1993). The accuracy of such estimates depends on the frequency of sampling and the reliability of the hydrological measurements. The modelling approach is, potentially, more powerful but requires detailed information on the hydrology of the catchment and the nature of the soil. Moore et al. (Chapter 11, this volume) describe the application of a simplified model (GWLF) which combines a realistic hydrology with an export coefficient procedure for estimating the flux of nitrate. This model is a good example of the 'fast and frugal' approach recommended by Carpenter (2003) and has proved effective at simulating both the observed and the projected variations in the flux of nitrate at a number of CLIME sites. The statistical approach is less sophisticated but can be used in

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situations where there are not enough data to support either budget calculations or modelling studies.

In this chapter, we use a statistical approach to investigate the effects of long-term changes in the weather on the supply and recycling of nitrate in lakes located in two, very different, climatic regions. Since the supply of nitrate to most of the selected lakes increased during the period of investigation, the raw time-series have been de-trended to focus on the effects associated with short-term (inter-annual) changes in the weather. The chapter includes an analysis of the large-scale climatic features that influence these variations and discusses the practical implications of the patterns observed in the two regions.

10.2 The Sites Included in the Analyses

Table 10.1 summarises the physical and chemical characteristics of the eight lakes included in the nitrate study. The sites have been arranged in increasing order of latitude from the north of the UK at 54°2'N to southern Finland at 61°0'N. All the sites were situated in lowland areas where the main source of nitrate was the drainage from the surrounding land. The smallest lake was Blelham Tarn in the UK, a productive lake with a surface area of only 0.1 km^2. The largest lake was Võrtsjärv in southern Estonia with a surface area of 270 km^2. The lakes with the highest proportion of cultivated land in their catchment were Blelham Tarn and Esthwaite Water in the UK and Võrtsjärv in Estonia. The only lakes with a significant loading of nitrate from point sources were Esthwaite Water and the South Basin of Windermere.

Table 10.1 The physical and chemical characteristics of the lakes included in the nitrate study

Site	Latitude (°N)	Altitude (m)	Area (km^2)	mean depth (m)	Residence time (years)	Arable land (%)	Nitrate (μg L^{-1})
Blelham Tarn	54.2	42	0.1	6.8	0.1	49	290–1,200
Esthwaite Water	54.2	65	1.0	6.4	0.2	44	280–800
Windermere (South Basin)	54.2	40	6.7	16.8	0.3	24	250–630
Windermere (North Basin)	54.4	40	8.0	25.1	0.5	22	240–620
Võrtsjärv	58.1	34	270	2.8	1.0	42	50–1,200
Mälaren (Galten)	59.4	1	61.5	3.4	0.1	9	5–1,250
Erken	59.8	11	23.7	9.0	7.4	10	0–2,050
Pääjärvi	61.0	103	13.4	14.4	3.3	17	480–1,100

10.3 Sampling Methods and Statistical Analysis

Samples of water for chemical and biological analyses were, typically, collected from the lakes at weekly or fortnightly intervals but some sampling dates were missed during the winter. In the English Lake District, the samples were collected with an integrating sampler (Lund and Talling, 1957). Elsewhere, samples were collected with closing bottles and weighted averages calculated for the water column. The time-series analyzed extended from January 1960 to December 2000, a period which included a number of particularly cold winters and very warm summers. Historically, a variety of statistical methods have been used to explore the factors responsible for the inter-annual variations in the concentration of nitrate. These include the analysis of variance (Casey and Clarke, 1979), autoregressive modelling (Naden and McDonald, 1989) and the analysis of the residuals in linear regressions (Worrall and Burt, 1998). Here, we used an adaptation of the method described by Worrall and Burt. In this procedure, the time-series are assumed to consist of two additive elements: the long-term trend and the short-term variation. Trend lines were fitted by least squares regression and the residuals used as a measure of the 'weather-related' variability. Tests with a variety of models showed that linear equations provided the best fit to the available data. The residuals in the regressions were checked for serial correlation and the results graphed to identify any systematic patterns or extreme values. Only one significant autocorrelation was detected in the data analyzed, a result that was assumed to be due to chance. Summary statistics were produced for each variable and the skewness and kurtosis values checked for agreement with a normal distribution. In almost every case, these parameters were within the range expected for a normal distribution. The only exception was the skewness value for the snow depth at Galten, a catchment that covers a very large part of southern Sweden.

10.4 The Seasonal Variations in the Concentration of Nitrate

In lakes, the seasonal variation in the concentration of nitrate is known to be influenced by a number of factors. These include: 1. Deposition from the atmosphere. 2. Flow pathways in the soil. 3. Chemical and biological processes in the soil. 4. The residence time of the lake. 5. The mixing characteristics of the lake. 6. The growth of phytoplankton. In the CLIME lakes, the factors that had the most pronounced effect on the dynamics of nitrate were the rate of supply from the surrounding land and the amount assimilated by the growth of phytoplankton (Chapters 3 and 14, this volume). All the sites were situated in areas where atmospheric deposition of nitrate accounted for a small proportion of the measured inputs. In each case, the seasonal variation in the concentration of nitrate followed a sinusoidal pattern with high concentrations being recorded in winter and low concentrations in late summer. The magnitude of the winter maximum was largely controlled by the transfer of nitrate from the surrounding land but that of the summer minimum was influenced by a variety of 'in lake' processes.

The basic pattern can be illustrated by comparing the time-averaged variation in the concentrations of nitrate in Blelham Tarn, Esthwaite Water and the North Basin of Windermere (Fig. 10.1a). These averages are based on samples collected every week between 1960 and 1978, every week in summer and every fortnight in winter between 1979 and 1991 and every fortnight after 1992. The lowest concentrations of nitrate were recorded in the deep North Basin of Windermere and the highest in Blelham Tarn, a shallow lake surrounded by agricultural land.

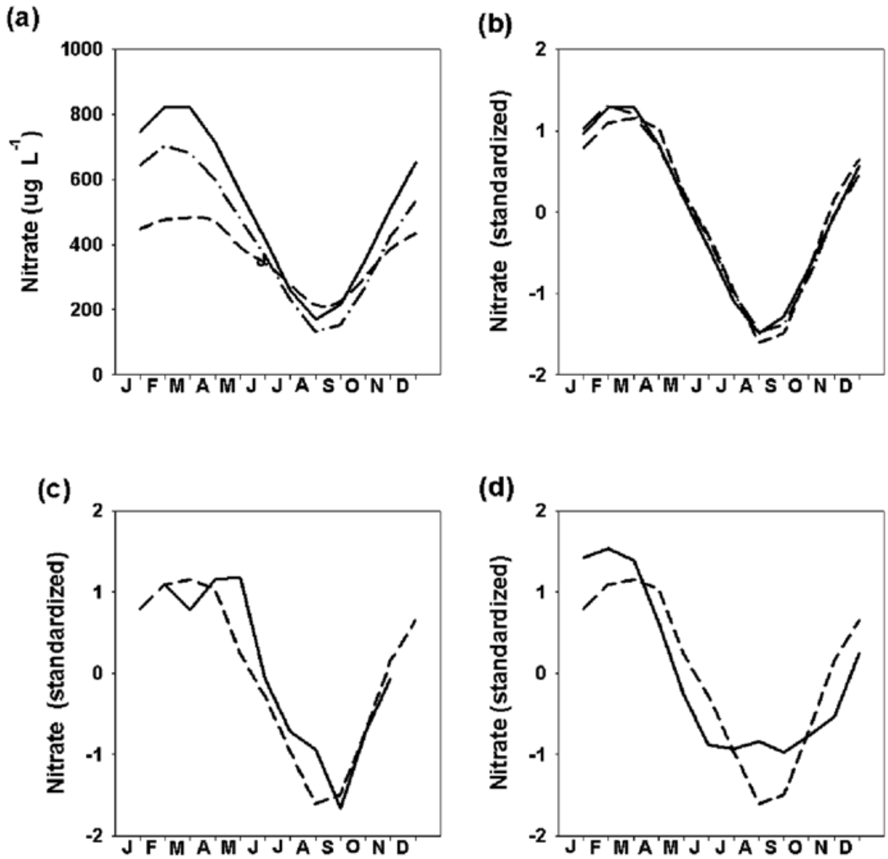

Fig. 10.1 (**a**) The seasonal variation in the concentration of nitrate in three English Lakes. (- - -) = Blelham Tarn, (-.-) = Esthwaite Water and (—) = the North Basin of Windermere. (**b**) The standardised concentrations of nitrate in the three English Lakes. (**c**) The standarised concentration of nitrate in Pääjärvi (—) compared with the North Basin of Windermere (- - -). (**d**) The standardised concentration of nitrate in Võrtsjärv (—) compared with the North Basin of Windermere (- - -)

Despite the trophic differences between the basins, the seasonal variations followed the same pattern in the three lakes. When the results were standardized, by subtracting the mean from each measurement and dividing by the standard deviation (Fig. 10.1b), the resulting curves were almost indistinguishable. When this procedure was used to compare the seasonal cycles observed in Pääjärvi and Võrtsjärv with the Windermere data, the between-site differences were more pronounced. In Pääjärvi (Fig. 10.1c), the seasonal cycle was quite similar to that observed in Windermere but both the winter maximum and the summer minimum appeared later in the year. In Võrtsjärv (Fig. 10.1d), the timing of the seasonal cycle was very close to that observed in Windermere but the winter concentrations of nitrate were higher and more nitrate remained in the open water at the end of summer. Relatively little is known about the seasonal dynamics of nitrate in this shallow lake but substantial amounts of nitrate are thought to be released from the shallow sediment during windy periods. Changes in the water level are known to be important (Nõges et al., 2003) and there has been a recent decline in the supply of nitrate from agricultural sources (Nõges et al., 2007).

The most striking feature of the standardized curves is the similarity of the seasonal cycles. At all sites, there is a well defined winter maximum followed by a progressive decline to a summer minimum. In the sections that follow, we demonstrate that the year-to-year variations in the magnitude of these maxima and minima provide the key to understanding both the direct and the indirect effect of the weather on the seasonal dynamics of nitrate. To simplify the comparisons between the different sites, we use the average March concentrations of nitrate as a measure of the winter maximum and the average September concentrations as a measure of the summer minimum.

10.5 The Long-Term Trend in the Winter and Summer Concentrations of Nitrate

All the lakes included in this study were situated in lowlands areas where there had been a progressive increase in the supply of nitrogen from agricultural land (Vinten and Smith, 1993; Burt and Johnes, 1997). The simplest measure of this long-term trend is the change over time of the nitrate concentrations measured in late winter. At that time of year, the supply rate is high and there is very little biological uptake in the receiving waters. Here, we used the average of all nitrate measurements taken in March as a measure of this 'winter' supply. The results in Table 10.2 show that the March concentrations measured in all the lakes increased during the period of study. This increase was statistically significant in seven out of the eight sites ($r > 0.40$, $p < 0.01$). The only exception was the Galten basin of Lake Mälaren where there was a small reduction ($r = -0.31$, $p < 0.10$).

Table 10.2 The long-term trend in the March ('winter') and September ('summer') concentrations of nitrate in eight CLIME lakes. The correlations are those calculated between the average March and September concentrations and the year of observation

Site	'Winter' trend (correlation)	'Winter' rate of change (ug L^{-1} Yr^{-1})	'Summer' trend (correlation)	'Summer' rate of change (ug L^{-1} Yr^{-1})
Blelham Tarn	0.42 (p < 0.01)	12.2	0.47 (p < 0.01)	4.3
Esthwaite Water	0.52 (p < 0.001)	10.1	0.24 (p < 0.10)	1.9
Windermere (S)	0.75 (p < 0.001)	6.7	0.28 (p < 0.10)	2.7
Windermere (N)	0.59 (p < 0.001)	5.3	0.16 (N.S.)	1.1
Võrtsjärv	0.58 (p < 0.01)	4.0	0.01 (N.S.)	0.1
Galten	−0.31 (p < 0.10)	−6.9	−0.66 (p < 0.001)	−18.8
Erken	0.52 (p < 0.001)	19.4	0.63 (p < 0.001)	20.9
Pääjärvi	0.84 (p < 0.001)	11.2	0.93 (p < 0.001)	32.0

The factors influencing the late summer concentration of nitrate are more complex and vary from lake to lake. Periods of heavy rain can transfer substantial amounts of nitrate from the surrounding land but the concentrations measured in the lake are usually determined by the amount assimilated by the phytoplankton in any particular year. Significant increases in the late summer concentration of nitrate were, however, recorded in five out of the eight lakes (Table 10.2), most notably the lakes located in Northern Europe.

Long-term trends of this kind greatly complicate the task of quantifying the impact of the inter-annual variations in the weather. Clear climatic effects can, however, be identified once the raw time-series have been de-trended to minimise these anthropogenic impacts. The examples in Fig. 10.2 shows how linear regressions were used to de-trend the nitrate data acquired for Blelham Tarn and the North Basin of Windermere. The results for Blelham Tarn (Fig. 10.2a) show that the March

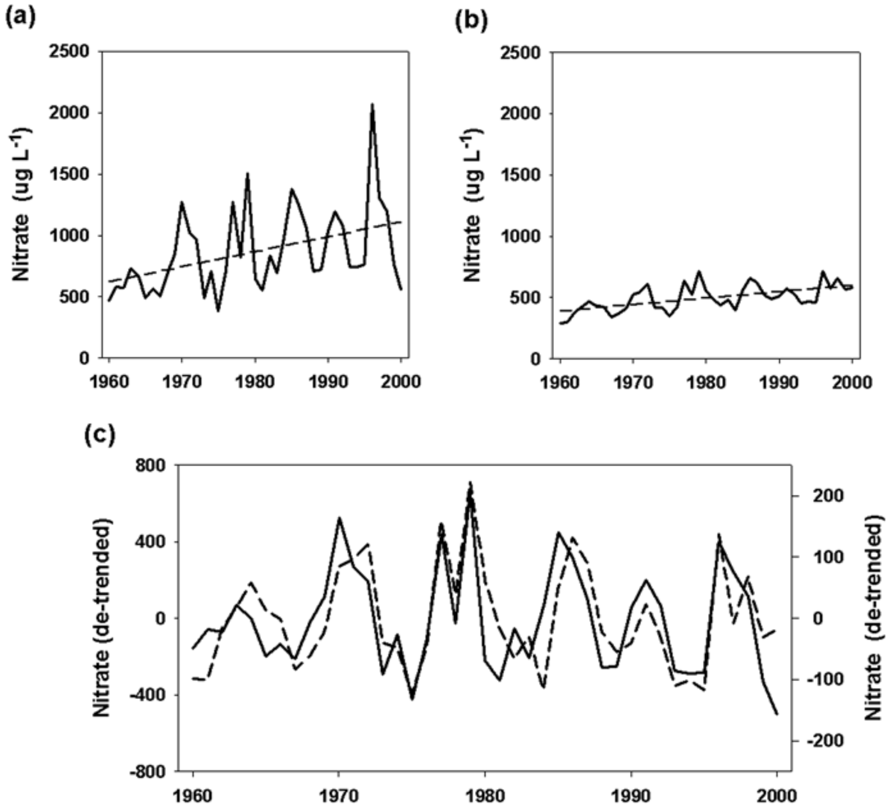

Fig. 10.2 (**a**) The year-to-year variations in the March concentration of nitrate in Blelham Tarn. (**b**) The year-to-year variations in the March concentration of nitrate in the North Basin of Windermere. (**c**) The year-to-year variations in the de-trended March concentrations of nitrate in Blelham Tarn (—) and the North Basin of Windermere (- - -)

concentration of nitrate increased from an average of 612 ug L^{-1} in the 1960s to an average of 1090 ug L^{-1} in the 1990s. The trend for the North Basin of Windermere (Fig. 10.2b) was less pronounced, but the values still increased from an average of 385 ug L^{-1} in the 1960s to an average of 552 ug L^{-1} in the 1990s. The fitted regressions showed that the trend in Blelham Tarn accounted for 18% of the observed variation whilst that for Windermere accounted for 35%. Figure 10.2c compares the de-trended time-series for the two lakes. Once the progressive effects of enrichment had been removed, there was a high level of temporal coherence between the two time-series (r = 0.75, $p < 0.001$). These results suggest that the factors responsible for the inter-annual variations in the de-trended winter concentrations of nitrate were associated with some fundamental process in the surrounding catchments i.e. the concentrations were not influenced by the size and trophic status of the individual lakes. Very similar 'generic' patterns were identified in the September time-series but only after these measurements had been adjusted to account for the nitrate assimilated by the phytoplankton during the summer (see Section 10.5.2).

10.5.1 The Impact of Year-to-Year Variations in the Winter Weather on the March Concentration of Nitrate

In this section, we use correlation analyses to identify the climatic factors that had the most significant effect on the winter concentration of nitrate in the eight CLIME lakes. At each site, the de-trended time-series were correlated with a range of meteorological variables and the most statistically significant variable assumed to be the key climatic driver.

In the North Basin of Windermere (Fig. 10.3a) the only significant factor was the average air temperature in December, January and February (r = -0.52, $p < 0.001$). Very similar correlations were observed in the other Western European lakes (Fig. 10.3b) where the coefficients ranged from a minimum -0.49 for the South Basin of Windermere to a maximum of -0.54 for Blelham Tarn. The most likely explanation for these negative relationships is the effect that mild winters have on the assimilation of nitrate in the surrounding soils. The factors responsible for regulating the supply of nitrate to the English Lakes is still a matter of some debate (George et al., 2004a; Tipping et al., 2007). Most of these processes are temperature-dependent but some are also influenced by the moisture content of the soil.

The results for Pääjärvi (Fig. 10.3c) showed that the factor that had the most significant correlation with the March concentration of nitrate was the depth of snow in the surrounding catchment (r = -0.74, $p < 0.001$). Very similar negative correlations were observed in the other Northern lakes (Fig. 10.3d), but these coefficients were not statistically significant. In these frozen lakes, the supply of nutrients is strongly influenced by the hydrological changes that follow the melting of snow (Arheimer et al., 1996; Rankinen et al., 2003). A key factor influencing the concentration of nitrate in this melt water is the depth of snow in the catchment. A deep covering of snow can insulate the soil to such an extent that microbes become active and there is

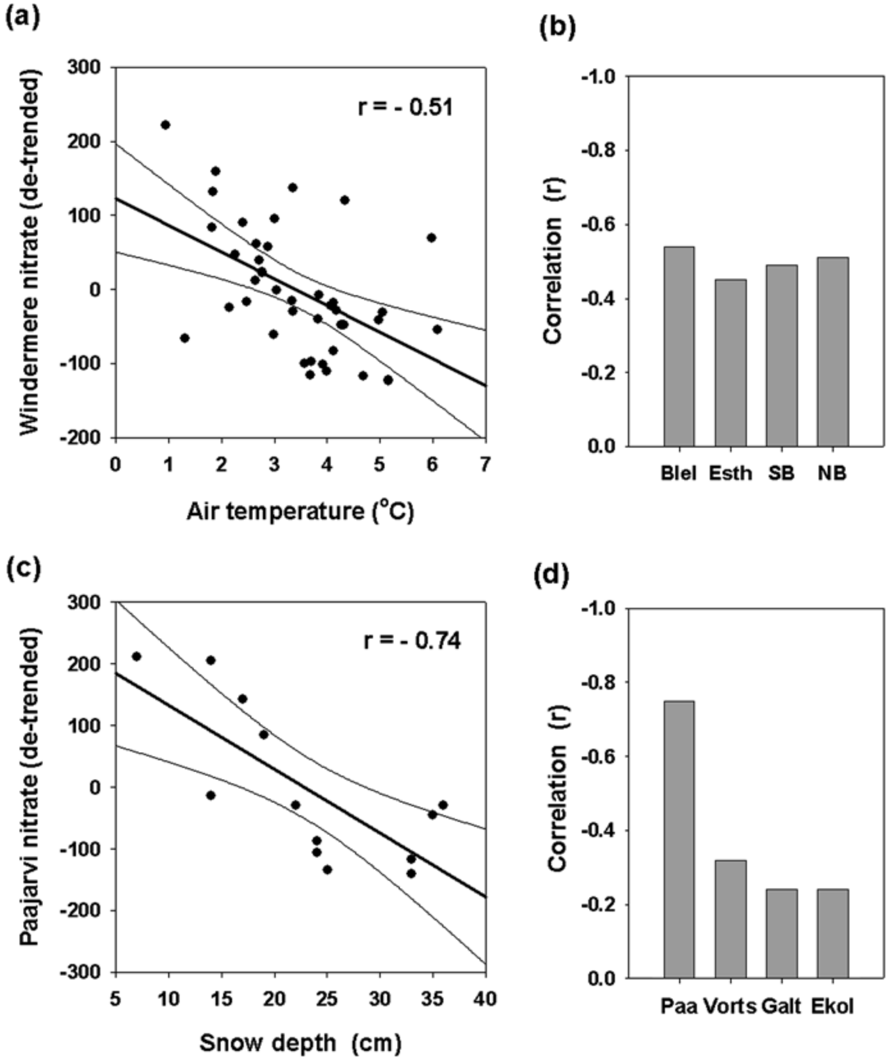

Fig. 10.3 (**a**) The relationship between the de-trended March concentration of nitrate in the North Basin of Windermere and the winter air temperature. (**b**). The correlation between the de-trended March concentration of nitrate and the winter air temperature in the four Western lakes. (**c**) The relationship between the de-trended March concentration of nitrate in Pääjärvi and the depth of snow in the catchment. (**d**) The correlations between the de-trended March concentration of nitrate and the depth of snow at the four Northern sites (the labels below the bars identify the different lakes)

even some general biological uptake (Schimel et al., 2004). In contrast, a light covering of snow exposes the soil to more frequent freeze/thaw cycles (Edwards and Cresser, 1992) which lead to the release of nutrients by root and microbial mortality (Boutin and Robitaille, 1994).

10.5.2 The Impact of Year-to-Year Variations in the Summer Weather on the September Concentration of Nitrate

In the CLIME lakes, the lowest concentration of nitrate was always recorded in September i.e. at the end of the growing season. One of the most important factors influencing the summer concentration of nitrate in all eight lakes was the amount assimilated by the phytoplankton. Long-term records of the seasonal variation in the biomass of phytoplankton were only available for some of these sites. The most complete records were those available for the English lakes and for Pääjärvi in southern Finland. Figure 10.4 shows the extent to which the summer biomass of phytoplankton at two of these sites influenced the September concentrations of nitrate. In Esthwaite Water (Fig. 10.4a), the inter-annual variations in the average summer (June to August) chlorophyll accounted for 18% of the recorded variation in the de-trended September nitrate (r = -0.42, $p < 0.01$). In Pääjärvi (Fig. 10.4b), the proportion accounted for was 32% (r = -0.57, $p < 0.001$).

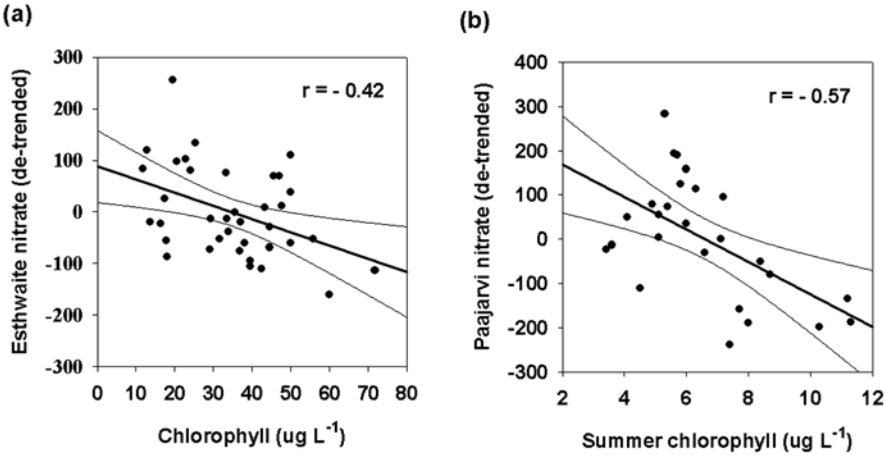

Fig. 10.4 The relationship between the de-trended September concentration of nitrate and the average summer biomass of phytoplankton in (**a**) Esthwaite Water and (**b**) Pääjärvi

In the analysis of the September time-series, we accounted for this assimilation by using the average summer biomass of phytoplankton as the dependent variable in a second de-trending regression. The residuals from this regression were then used as a measure of the assumed effect of the observed variations in the weather. The summer biomass was the average of measurements taken in June, July and August i.e the same averaging period used for the meteorological variables. The results showed that the only meteorological variable that had a significant effect on the residual September nitrate was the summer wind speed. Figure 10.5 shows the extent to which the residual concentrations of nitrate in Esthwaite Water and Pääjärvi were correlated with the average summer wind speed. In Esthwaite Water (Fig. 10.5a) the year-to-year variations in the wind speed accounted for 40% of the observed

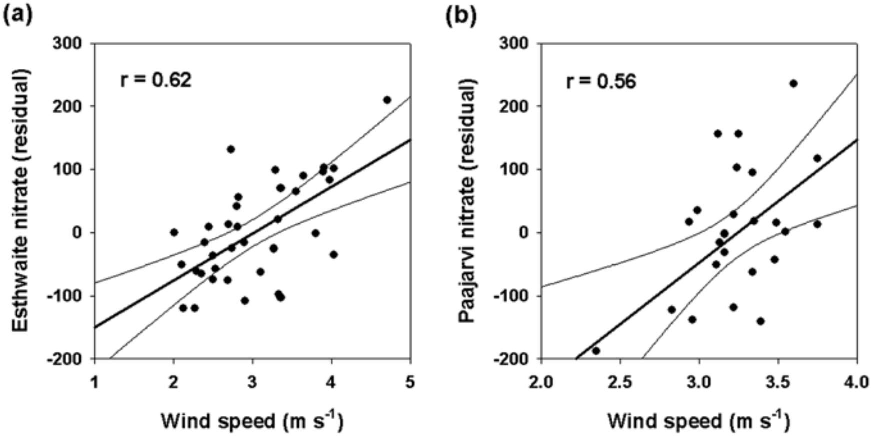

Fig. 10.5 The relationship between the residual September concentration of nitrate and the average wind speed in June, July and August in (**a**) Esthwaite Water and (**b**) Pääjärvi

inter-annual variation ($r = 0.62$, $p < 0.001$) in the residual nitrate. In Pääjärvi (Fig. 10.5b), the impact of the wind was less pronounced but the fitted regression still explained 30% of the variance in the de-trended time-series ($r = 0.56$, $p < 0.001$). The most likely explanation for this effect is the impact that changes in the wind speed have on the transfer of nutrients from deep water into the epilimnion. In Esthwaite Water, most of the nitrate in the epilimnion is assimilated by the growth of phytoplankton but substantial concentrations still remain in the metalimnion (Heaney et al., 1986). In Pääjärvi, the vertical variations in the distribution of nitrate are less pronounced but the concentrations measured in the upper layers of the hypolimnion are always higher than those measured in the epilimnion.

10.6 The Impact of Large-Scale Changes in the Atmosphere on the Flux of Nitrate

The meteorological variables that influence the inter-annual variations observed in these lakes are regional manifestations of much larger-scale features in the atmosphere. Climatologists have devised a number of different methods to quantify these variations, some of which are described in Chapters 16 and 17 of this volume. In winter, the atmospheric feature that has the most pronounced effect on the dynamics of lakes in Northern, Western and Central Europe is the North Atlantic Oscillation (Straile et al., 2003). A common measure of the North Atlantic Oscillation (NAO) is the index described by Hurrell (1995). This is based on the pressure gradient that develops between Stykkisholmur in Iceland and Lisbon in Portugal. Positive values of the index are associated with mild, wet winters and negative values with colder, drier conditions.

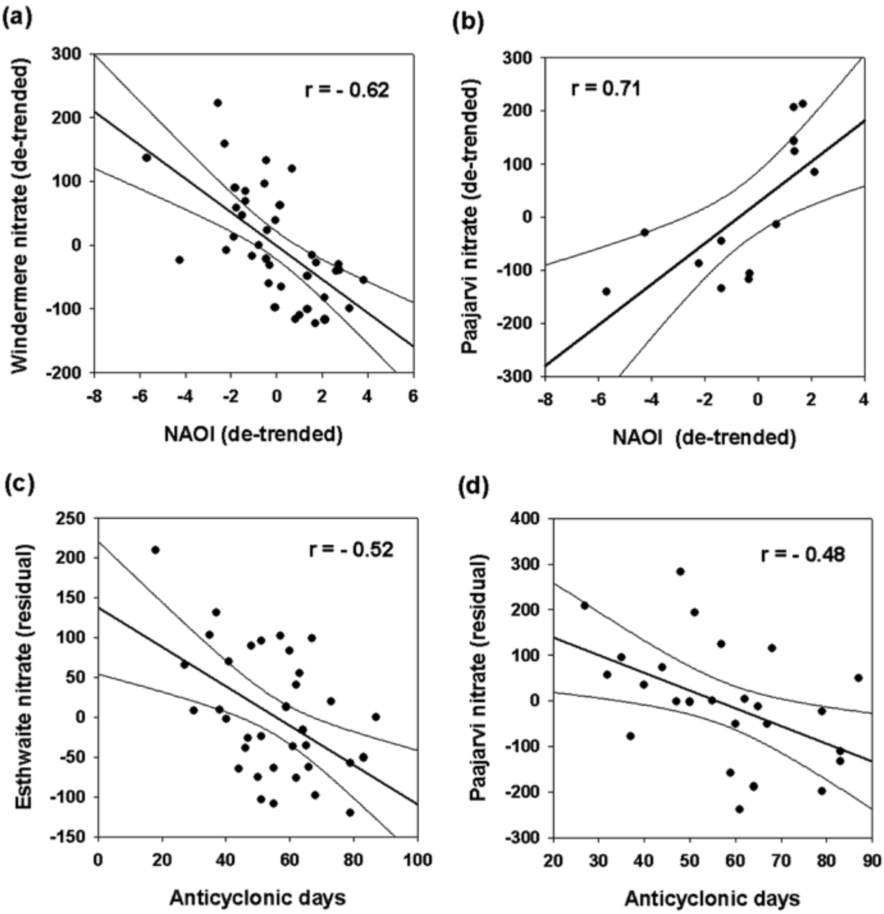

Fig. 10.6 (**a**) The relationship between the de-trended March concentration of nitrate in the North Basin of Windermere and the de-trended North Atlantic Oscillation Index. (**b**) The relationship between the de-trended March concentration of nitrate in Pääjärvi and the de-trended North Atlantic Oscillation Index. (**c**) The relationship between the residual September concentration of nitrate in Esthwaite Water and the number of anticyclonic days. (**d**) The relationship between the residual September concentration of nitrate in Pääjärvi and the number of anticyclonic days. The anticyclonic frequencies are based on the weather typing procedure described by Lamb (1950)

Figure 10.6a shows the relationship between the de-trended March concentration of nitrate in the North Basin of Windermere and the de-trended North Atlantic Oscillation Index (NAOI). The relationship is strongly negative and the fitted regression explained 39% of the observed variation (r = -0.62, p < 0.001). In Pääjärvi (Fig. 10.6b) the variations in the NAO had a very different effect on the winter flux of nitrate (George et al., 2004b). Here, the relationship between the de-trended March concentration of nitrate and the de-trended NAOI was positive and the fitted regression explained 49% of the observed variation (r = 0.70, p < 0.001). These differences can be explained by the different factors regulating the supply of nitrate at the two sites. In Windermere, the critical factor is the enhanced terrestrial uptake of nitrate in mild winters, when the NAOI is strongly positive. In Pääjärvi, the most influential climatic factor is the depth of snow. In negative NAO years, the snow tends to be deeper, and insulates the soil to such an extent that more nitrate is assimilated during the winter.

The factors that affect the residual concentration of nitrate in late summer are more complex and reflect both changes in the external supply and variations in the internal recycling of the nutrient. In the sites analysed here, the only significant meteorological factor was the wind speed, which regulated the vertical transport of nitrate in the water column. Recent high-resolution measurements of the vertical variations in the temperature of lakes in the Atlantic region (George unpublished) have shown that these mixing events can be very short and strongly influenced by day-to-day variations in the synoptic situation. One way of characterising these variations is the 'weather typing' approach described by Lamb (1950). The original Lamb system was based on the subjective analysis of daily weather maps but the current system is based on the automated procedures described by Jenkinson and Collison (1977). In Northern and Western Europe, the synoptic situation that has the most pronounced effect on the mixing characteristics of the lakes is the anticyclonic type which is characterised by high pressure and low wind speeds (Chapter 16, this volume).

Figure 10.6c shows the relationship between the residual September concentration of nitrate in Esthwaite Water and the number of anticyclonic days recorded during the summer. The results show that there is a strong negative relationship between the two variables and the fitted regression explained 26% of the observed variation (r = -0.52, $p < 0.001$). The same general effect was observed in Pääjärvi (Fig. 10.6d) where the fitted regression explained 23% of observed variation (r = -0.48, $p < 0.05$). A surprising feature of these results is the strength of the correlation observed at the Finnish site. Pääjärvi lies outside the area covered by the Lamb system of weather typing but the wind speeds recorded there were influenced by the pressure patterns that developed over the British Isles.

10.7 Discussion

The concentration of nitrate in most European waters has increased in recent years (Heathwaite et al., 1993). This increase has given rise to a number of problems, such as the acidification of lakes (Tipping et al., 1998) and changes in the qualitative composition of phytoplankton (Ferber et al., 2004; Weyhenmeyer et al., 2007). The factors that influence the atmospheric deposition of nitrate and its transformations in the soil have been the subject of particularly intensive study (Fowler et al., 1989; Roberts, 1990; Wayne, 1993; Monteith et al., 2000; Monteith et al., 2001). Less attention has hitherto been paid to the effects of year-to-year changes in the weather on the transfer of nitrate from terrestrial to aquatic systems.

The processes that regulate the transfer of nitrate to surface waters are very different from those that control the supply of phosphate. Nitrate ions are not so tightly bound to the soil matrix and are more readily mobilised by rain. The relationship between the quantity of nitrate leached from the catchment and that measured in the lake is, however, quite complex. It can be influenced by the depth of a lake, its residence time and the growth of different functional groups of phytoplankton. In this chapter, we have used a statistical approach to investigate the effects of long-term changes in the weather on the supply and recycling of nitrate in a number of very different lakes. Long-term records of the seasonal variations in nitrate are still rare in European lakes. At most sites, samples are either collected at infrequent intervals or measurements confined to a particular time of year. In the lakes analysed here, samples were collected at sufficiently frequent intervals to quantify the impact of long-term changes in the weather on the measured concentrations of nitrate.

In the eight lakes selected for study, the seasonal variations in the concentration of nitrate followed the same general pattern with the highest concentrations recorded in winter and the lowest in the summer. The main factor regulating the winter maximum was the quantity of nitrate leached from the surrounding land. The factors influencing the summer minimum were more complex and included the mixing characteristics of the lakes and their trophic status. Significant increases in both the winter and summer concentrations of nitrate were recorded in six out of the eight lakes with the most pronounced increases being recorded during the winter. These increases are, however, modest when compared to those reported elsewhere (UNEP/WHO, 1988; Heathwaite and Burt, 1991). The trends, nevertheless, meant that all the time-series had to be de-trended before the key climatic drivers could be identified. There is consequently a risk that this procedure will have obscured some climatic effects, such as those associated with the slow, progressive change in the climate. Results presented elsewhere (e.g. George et al., 2007; Nõges et al., 2007) suggest that these progressive changes are, however, less important than the inter-annual variations analyzed here.

In the Western lakes, the factors responsible for the observed variation in the winter concentration of nitrate are still a matter of debate. Monteith et al. (2000) suggested that the year-to-year variations in the nitrate content of their upland lakes were the result of the physical effect of frost on the leaching properties of the soil. In contrast, George et al. (2004a), working in the English Lake District, assumed that the factor responsible was the increased terrestrial uptake of nitrate in mild winters. The uptake of nitrate in soil is mediated by two different processes: assimilation by vascular plants and bacterial de-nitrification. The first mechanism is known to be important in summer but plants can also assimilate substantial quantities of nitrate when the soil is cold (Lain et al., 1994). Microbial de-nitrification in soil has long been regarded as an important route of nitrate removal (Ryden and Rolston, 1983; Lensi et al., 1991). At some sites the maximum rates of de-nitrification have been recorded in the autumn (Pinay et al., 1993) whilst at others the highest rates have been recorded during the winter (Groffman and Hanson, 1997). Very little is known about the rates of de-nitrification in the soils of the Lake District. The land is, however, very wet and poorly drained soils are known to have a high de-nitrification potential (Davidson and Swank, 1987; Groffman and Tiedje, 1991). The absolute amount of nitrate leached from these catchments may also be affected by soil type. In a recent study of two lakes in the west of the English Lake District, Tipping et al. (2007) report that the highest concentrations of nitrate were found in samples collected from areas dominated by ranker soils. In contrast, no significant nitrate leaching was recorded in the sub-catchments dominated by brown podzols, another common soil type.

In the Northern lakes, the factors regulating the winter dynamics of nitrate have been the subject of particularly intensive study. The controlling factors are believed to be the depth of snow, the timing of snow melt and the hydrological characteristics of the catchment. Most researchers report a net mineralization of nitrogen under snow but there is also evidence of some seasonal immobilisation in a few soil types (Clein and Schimel, 1995; Schmidt et al., 1999). Snow is an effective insulator and the temperature of a soil covered by deep snow can be more than 20°C higher than an adjoining area covered by little snow (Edwards and Cresser, 1992; Schimel et al., 2004). There are two possible explanations for the inter-annual variations observed in Pääjärvi. The first is that the alternate periods of freezing and thawing, a feature of mild winters, result in a net increase in the leaching of nitrate. If this hypothesis is correct, then the mechanism regulating the winter supply of nitrate in Pääjärvi is the same as that suggested by Monteith et al. (2000) for upland lakes in the UK. The second is that a deep covering of snow in cold winters, insulates the soil to such an extent that significant amounts of nitrate are assimilated by soil microorganisms. In this case, the mechanism is the same as that suggested by George et al. (2004a) for the English lakes. Regional comparisons of this kind suggest that more attention should be paid to the processes that influence the flux of nitrate during the coldest months in the year. Rankinen et al. (2003) have shown that these 'cold processes' are important in northern catchments and should be included in any models designed to simulate the seasonal flux of nitrate. When their INCA model was used to simulate the flux of nitrate in the Simajoki river basin in southern Finland, it only produced realistic results when the processes operating at sub-zero temperatures were included.

The factors influencing the summer concentration of nitrate include a number of biological as well as physical processes. Nitrate usually plays a secondary role in the productivity of lakes but the assumption that the element is never limiting is an oversimplification (Heathwaite et al., 1993). Sustained periods of nitrogen limitation are rare, but short-lived shifts in the N to P ratio can have an important effect on the seasonal dynamics of many species of phytoplankton (Tilman et al., 1982; Smith, 1983). In the lakes analysed here, we observed a progressive decline in the eplimnetic concentrations of nitrate as the growing season progressed. In the English lakes, the average proportion lost was 64% and the individual estimates ranged from a minimum of 52% in the North Basin of Windermere to a maximum of 78% in Esthwaite Water. The main factor responsible for this decline was the assimilation of nitrate by the phytoplankton. Once the September measurements had been detrended to account for this effect, significant climatic effects could be detected in the two lakes selected for detailed study. Weyhenmeyer et al. (2007) report that the proportion of nitrate lost over the growing season was very similar in a large sample of European lakes. Their suggestion that periods of nitrogen limitation are becoming increasingly common is not substantiated by our observations in the English Lakes. For example, in Esthwaite Water, the most productive lake, nitrate concentration below 20 ug L^{-1} were recorded on two occasions in the 1960s, fifteen occasions in the 1970s, two occasions in the 1980s and four occasions in the 1990s. The most striking feature of these analyses was the effect that the observed variation in the weather had on the residual September concentration of nitrate. The most significant correlation was that observed with the average summer wind speed. Related variables, like air temperature and solar radiation, had no direct effect, even though they have an indirect effect on the intensity of wind-induced mixing.

The historical analyses summarized here represent the first steps in a study that also included the model simulations described in Chapter 11. These simulations support most of the conclusions drawn here and confirm that the factors regulating the flux of nitrate were different in the two climatic regions. At the Northern sites, the key factor was the change in the timing of the spring thaw. This event has a major effect on the seasonal dynamics of nitrate and is, in turn, regulated by the temperature of the soil as well as of the air. At the Western sites, the most important factor was the projected increase in the winter rainfall. This may lead to a progressive increase in the winter load of nitrate but this trend could be reversed if increasing amounts of nitrate are retained by biological uptake in the catchment.

Historical analyses, of this kind, provide a useful perspective on the ‘nitrate problem’ as currently conceived. The legislation that governs the permitted concentration of nitrate in European waters was introduced at a time when climate change was not part of the political agenda. The results presented here demonstrate that changes in the climate can influence the supply and re-cycling of nitrate in a number of ways. Most of the lakes covered by our studies were relatively deep and were situated in areas where the atmospheric deposition of nitrate accounted for a small component of the annual inputs. Very different responses are to be expected in shallow lakes and in lakes where atmospheric deposition of nitrogen has a significant effect on their chemistry and biology (Evans and Monteith, 2001). More effort should now be devoted to exploring the factors that regulate the flux of nitrate in lakes that are likely to experience a very rapid change in the climate (Weyhenmeyer et al., 2004). Some attention also needs to be given to the practical consequences of the different seasonal patterns that will emerge as the world becomes warmer. Lakes are widely regarded as natural ‘filters’ that limit the ecological damage caused by diffuse inputs of nutrients. There is now no doubt that the efficiency of these ‘filters’ will change as the world becomes warmer. At some sites, these changes may have little effect on the ecological status of the lake. At others, the criteria used to define the status of the lake may have to change and new strategies devised to manage the soils and the drainage in the surrounding land.

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Chapter 11
Modelling the Effects of Climate Change on the Supply of Inorganic Nitrogen

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11.1 Introduction

Human-induced changes in the nitrogen cycle due to the increased use of artificial fertilisers, the cultivation of nitrogen-fixing crops and atmospheric deposition have made nitrogen pollution to surface waters a long-standing cause for concern. In Europe, legislation has been introduced to minimise the risk of water quality degradation from excessive nitrogen inputs e.g., the European Union Nitrates Directive (EU, 1991), Drinking Water Directive (EU, 1998) and Water Framework Directive (EU, 2000). Coastal regions in particular have been an important focus, since coastal eutrophication has been attributed to increased fluxes of nitrogen from the landscape (Howarth et al., 1996; Boesch et al., 2006). While nitrogen is typically not the limiting nutrient in inland waters, increases in the nitrogen supply to rivers and lakes can impact the N:P ratio, influence the structure of aquatic food webs (Elser and Urabe, 1999; Arbuckle and Downing, 2001), regulate the seasonal development of phytoplankton (Van den Brink et al., 1993) and promote the growth of aquatic macrophytes (Kirchmann et al., 2004).

The export of dissolved inorganic nitrogen (DIN) from catchments is influenced by a number of physical, chemical and biological factors. Catchment-related impacts include the spatial and temporal variability in catchment hydrology and land management. Climate-related impacts include the effects of temperature and soil moisture on biogeochemical processes and soil moisture levels and the effects of precipitation and snowmelt on the flushing of DIN from soils. Nitrate, the main form of DIN in soils, is highly mobile and is easily leached from the soil matrix (Carpenter et al., 1998). Consequently, the main export of DIN typically coincides with high flow events (e.g., David et al., 1997; George et al., 2004; Ocampo et al., 2006). Other factors influencing the leaching of nitrogen include soil type and topography, changes in the cropping system or animal husbandry and the application of fertilisers and manures in agricultural catchments (Hooda et al., 2000).

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Nitrogen losses from agriculture remain a significant source of nutrient loading to inland and coastal areas (Arheimer and Brandt, 2000; Andersson and Arheimer, 2003; Arheimer et al., 2005). Increases in nitrate levels have no single origin and a range of management strategies for reducing nitrate leaching are therefore required (Kirchmann et al., 2004).

In many forested catchments in Europe and North America, increased atmospheric deposition of nitrogen in excess of assimilation capacity has led to nitrogen saturation (Vitousek et al., 1997; Aber et al., 2002). Some recovery of streams and lakes affected by atmospheric deposition has been observed in recent years. For example, according to an analysis of acidification trends at 189 sites in Europe and North America by Skjelkvåle et al., 2005, recovery of many surface waters has been observed; however, this is largely attributed to reduced $SO_4{}^{2-}$ deposition rather than changes in NO^{3-} concentrations. In Finland, this recovery can primarily be attributed to decreased $SO_4{}^{2-}$ deposition relative to base cation deposition (Vuorenmaa and Forsius, 2008). Some observed variations in the deposition rate are difficult to interpret, since the trends are examined on a decadal scale and the recovery of the catchments may take several centuries (Skjelkvåle et al., 2005; Grimvall et al., 2000). These variations are also influenced by other factors such as disturbance events and climate variability (Aber et al., 2002). Where forest disturbance and age of the stand cannot account for decreases in nitrogen leaching, changes in microbial immobilisation or denitrification are plausible alternative explanations for the reduced rates of nitrate loss (Goodale et al., 2005). At present, catchment responses to nitrogen saturation and subsequent leakage are difficult to predict (Evans et al., 2006), and the conceptual framework required to understand and predict nitrate leaching is still being developed (e.g., Emmett, 2007).

In this chapter, we describe how we used an established catchment model to quantify the effects of both observed and projected changes in the climate on the loading of DIN to lakes and rivers in five European catchments. Increases in nitrogen loads are anticipated for regions where future climate projections are for warmer, wetter conditions (Fink and Kralisch, 2005; Jöborn et al., 2005 and Chapter 2, this volume). A related chapter (Chapter 10) explores the effect of year-to-year changes in the weather on the supply and recycling of nitrate in lakes and assesses their likely response to future changes in the climate.

The model used in our study is the Generalized Watershed Loading Functions model (GWLF). The basic structure of the model has already been described in Chapter 3, and Chapter 9 explains how it was used to simulate the effects of future changes in the climate on the supply of phosphate-phosphorus. Here, we use the same approach to estimate future DIN loads using inputs from the same Regional Climate Models perturbed by the same emissions scenarios (IPCC SRES A2 and B2). Details of the six climate scenarios used are given in Chapter 2 and summarized in Table 11.1. This chapter includes an assessment of the DIN model performance, an overview of the DIN loading projections and a discussion of the implications of the projected changes in the DIN flux for the seasonal dynamics of the selected lakes. The examples presented are taken from catchments in Northern and Western Europe and the lakes cover a range of different sizes and trophic state.

Table 11.1 Future climate scenarios used in the CLIME DIN simulations

Abbreviation	GCM	RCM	Greenhouse gas scenario
H A2	HadAM3h	RCAO	A2
H B2	HadAM3h	RCAO	B2
E A2	ECHAM4/OPYC3	RCAO	A2
E B2	ECHAM4/OPYC3	RCAO	B2
Had A2	HadAM3p	HadRM3p	A2
Had B2	HadAM3p	HadRM3p	B2

11.2 Modelling Approach

A number of simulation models are available for predicting nitrogen flux from catchments to receiving waters. These models differ in their complexity, and the diversity in these approaches reflects the purpose and scale of their intended application. Several process-based models were described and compared in the EUROHARP project and covered a range of scales from plots to large catchments (Schoumans and Silgram, 2003). Statistical models are commonly applied at broader regional to global scales (see Howarth et al., 1996; Smith et al., 1997; Caraco and Cole, 1999; Fink and Kralisch, 2005; Seitzinger et al., 2005). The choice of the most appropriate model depends primarily on the objectives of the study (Quinn, 2004). The model used in CLIME (GWLF) can be described as a model of moderate complexity (Haith et al., 1992). Some recent adaptations of the model include a stand-alone Windows application (Dai et al., 2000), a spreadsheet version (Hong and Swaney, 2007), a database version (Smedberg et al., 2006) and the dynamic approach used here (Schneiderman et al., 2002). Whether used in its original QBasic form (Haith et al., 1992) or the Vensim version used here (Ventana Systems, Inc.), GWLF offers the benefits of being relatively easy to apply to catchments that are very different in size and include mixed land uses. Land use and rainfall rates, two of the key influences on nitrogen export at the catchment scale, are both represented in the model (Quinn, 2004).

We chose GWLF as the preferred approach since the model has already been used with great success to test different nutrient management strategies in several US catchments (Haith and Shoemaker, 1987; Swaney et al., 1996; Schneiderman et al., 2002). The very large number of simulations required in CLIME demanded a model that was easy to drive with the projected changes in hydrology and required less detailed information on the transformation processes in the selected catchments. We do not dismiss the importance of process-based models, such as those described by Birkenshaw and Ewen (2000), Alexander et al. (2002) and Wade et al. (2002), but they require more data than was commonly available at the CLIME sites. Given the anticipated impact of the climate on temperature, precipitation, runoff and soil moisture we chose a model that focused on the transport term and assumed simple land use-specific concentrations to generate the associated nutrient loadings. This flow rate approach was considered particularly appropriate for dissolved inorganic nitrogen (DIN) given the influence of flushing rates and changes in hydrological pathways on export rates (Carpenter et al., 1998; Ocampo et al., 2006).

11.2.1 Key Features of the Model

In GWLF, simple loading functions for DIN coupled to dynamic changes in hydrology are used to estimate loads from the catchment. The model runs on a daily time step and the results are then aggregated for longer periods. The land-use specific nutrient concentrations used in our simulations are based on the best information available for each site. Basin-wide concentrations for baseflow are derived from water quality data averaged for stream discharge at or below the 20th percentile for the period of observation. The net DIN flux is a product of the combined contributions from rapid surface runoff and a slower baseflow, and the flow-weighted concentrations of these fractions. GWLF is sometimes referred to as a ‘semi-distributed’ model because the contributions from different land uses vary according to their runoff characteristics and associated nutrient concentrations. The contributions of NO_x from atmospheric deposition are implicit in the land use inputs and are not considered as a separate term in the model. Point source inputs were taken from historical records and the same values were used for the control and future climate simulations to give a ‘business as usual’ weighting to the loading estimates. Similarly, contributions from septic systems, agricultural practices, land cover and manure application were modelled for the historical period and held constant for the future climate simulations. Timing of manure spreading and fertiliser application can have profound effects on nitrogen inputs from arable land. Where information was available, these inputs were included in the land use-specific estimates of DIN concentrations. The site-specific loading functions for inorganic nitrogen concentrations used to calibrate the model were assigned to the land use classes listed in Table 11.2.

Climate impacts were modelled by driving the simulations with the down-scaled outputs of the Regional Climate Models described in Chapter 2. In our projections, synthetic time series of the daily weather representing 100 iterations of 30-year periods were used to drive both the control (1961–1990) and the future (2071–2100) simulations with GWLF. The synthetic time series were obtained from weather generator outputs for Ireland, the UK, and Estonia (see Chapter 2) and by an alternative statistical downscaling method for Finland and Sweden (see Hay et al., 2000 and Chapter 3). Adjustment for the changing length of the growing season was made using the method described by Mitchell and Hulme (2002). By adopting this stochastic approach, we were able to run the GWLF model of DIN flux for five sites in Western and Northern Europe and quantify the uncertainty associated with the calculated projections.

Table 11.2 The characteristics of the five CLIME sites used for the DIN simulations

Land use classification	Area (ha) by site				
CORINE class	Esthwaite (UK)	Flesk (IE)	Mustajoki (FI)	Arbogaån (SE)	Tarvastu (EE)
Continuous urban	0	1	0	58	0
Discontinuous urban	0	188	0	3,188	4
Industrial	0	0	0	958	0
Road and rail networks	0	170	10	168	18
Airports	0	0	0	84	0
Mineral extraction sites	0	0	0	348	1
Dump sites	0	0	0	200	0
Farmyards	0	380	0	12	1
Green urban areas	0	0	0	2,036	4
Sport and leisure facilities	0	65	0	814	2
Non-irrigated arable land	0	0	0	44,576	2,800
Fruit trees/berry plantations	0	0	0	0	20
High productivity pastures	867	2,681	0	0	0
Mixed productivity pastures	0	2,286	0	6,835	530
Low productivity pastures	0	5,170	0	0	820
Complex cultivation patterns	0	99	998.4	0	0
Principally agriculture	0	842	0	0	0
Broad-leaved forest	210	135	0	1,3446	1,800
Coniferous forest	315	1,658	5,145.6	174,426	1,300
Mixed forest	0	0	0	25,255	800
Natural grasslands	170	3,064	0	0	900
Moors and heathlands	0	5,622	0	0	0
Transitional woodland shrub	0	60	0	6,457	300
Bare rocks	0	0	0	2	0
Sparsely vegetated areas	0	348	0	0	0
Inland marshes	0	92	0	821	0
Exploited peat bogs	0	238	0	0	0
Unexploited peat bogs	0	9,092	1,536	14,053	200
Water courses	0	0	0	649	50
Water bodies	0	346	0	26,389	10
TOTAL AREA (ha)	1,562	32,532	7,690	380,775	9,560

11.2.2 Model Calibration and Validation

The structure and the performance of the hydrological component of the GWLF model have already been described in Chapter 3. The sites used for the DIN simulations were the Esthwaite Water catchment (United Kingdom, UK); the River Flesk sub-catchment of Lough Leane (Ireland, IE); the Mustajoki sub-catchment of Lake Pääjärvi (Finland, FI); the Arbogaån river sub-catchment of Galten basin, Lake Mälaren (Sweden, SE); and the Tarvastu sub-catchment of Lake Võrtsjärv (Estonia, EE). The sub-catchments modelled represent a significant portion of the catchment surrounding each lake as shown in Table 11.2. This table also shows the diversity

Table 11.3 Calibration of dissolved inorganic nitrogen (DIN) at five CLIME sites

Site	Calibration period	Sampling frequency	Observed DIN concentration range (mg L^{-1})	Modelled DIN concentration range (mg L^{-1})	DIN adjustment factor
Esthwaite (UK)	1997	monthly	0–1.25	0.11–1.37	0.89
Flesk (IE)	1999–2004	48 hr composite	0.07–1.82	0.24–0.76	0.89
Mustajoki (FI)	1994–2004	weekly	0.22–4.65	0.73–1.75	1.01
Arbogaån (SE)	1980–1991	monthly	0.07–1.09	0.39–0.57	1.7
Tarvastu (EE)	1990–2002	monthly	0.17–5.41	1.7–4.45	3.5

in the selected catchments, expressed in land cover types based on the CORINE classification system used in Europe (European Environment Agency, 2000).

We begin by showing how the model was calibrated for DIN (generally the sum of NO_3^--N, NO_2^--N, and NH_4^+-N). In most the NH_4^+ contribution is low. For example, in Mustajoki (FI) it accounts 4% of the total nitrogen (TN) load and 7% of DIN load and in Tarvastu (EE) it contributes 6% of TN load and 8% of DIN load. The calibration factors used to adjust model concentrations (the DIN adjustment factors in Table 11.3) were obtained by minimising the squared deviations between simulated and measured values in the time-series of flow weighted concentrations. The adjusted values provided a much closer match between measured and simulated DIN loads. Nash-Sutcliffe values (Nash and Sutcliffe, 1970) for DIN loads ranged between 0.31 and 0.81. The performance of the model for nutrient loading estimates was not as good as the hydrological model alone, because there are inherent limitations to using generalised values for DIN concentrations.

A fundamental characteristic of loading functions is that they are based on long-term average concentrations. Consequently, they underestimate the influence of any short-lived 'peaks' that appear in the time-series. By comparing the measured and modelled DIN loads, GWLF was able to capture the general dynamics of dissolved organic nitrogen flux at all sites listed in Table 11.2. At sites where samples were collected at frequent intervals (e.g., Flesk and Mustajoki) there was a very close correspondence between the measured and modelled DIN loads (Fig. 11.1a, b). Sites with less frequent sampling correctly reproduced the timing of peak DIN loads, but the magnitude of these peaks was sometimes underestimated. Despite these limitations, our results imply that the GWLF model can be used with some confidence to make DIN flux projections.

A comparison of measured and modelled daily loads under current climate conditions (Table 11.4) shows that there was a very good agreement between both the median values and the observed and calculated ranges. In our 'warm world' projections we used the same model parameters to estimate the DIN loads under future climatic conditions.

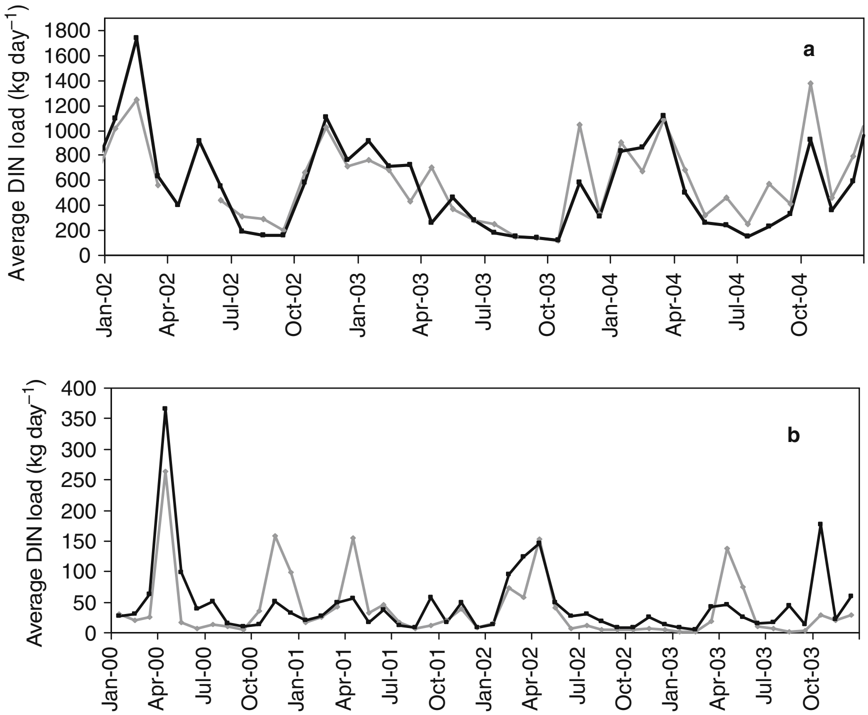

Fig. 11.1 Examples of DIN load estimation (measured = *grey*; modelled = *black*), (**a**). Flesk (IE), 2002–2004 (**b**). Mustajoki (FI), 2000–2003

Table 11.4 Measured and modelled dissolved inorganic nitrogen loads, kg day^{-1}

Site	Measured median daily load	Modelled median daily load	Measured load range	Modelled load range
Esthwaite (UK)	20	11	4–533	<1–863
Flesk (IE)	377	287	15–9,664	108–6,627
Mustajoki (FI)	15	26	1–778	4–915
Arbogaån (SE)	2,522	2,133	258–18,138	681–14,028
Tarvastu (EE)	84	126	3–910	37–1,965

11.3 The Projected Change in the Flux of Dissolved Inorganic Nitrogen (DIN)

Climate forcings for the CLIME project were taken from the Rossby Centre Atmosphere-Ocean (RCAO) regional climate model and the Hadley Centre regional climate model (HadRM3p) using boundary conditions from two general circulation models: the Max Planck Institute ECHAM4/OPYC3 and the Hadley Centre HadAM3H for two IPCC emissions scenarios (A2, B2). Further details of the climate scenarios are given in Chapter 2 and summarised in Table 11.1. The use of multiple models allows for some consideration of the range of uncertainty in the model projections. A 30-year period was used for a control (1961–1990) and future (2071–2100) simulations.

The method used to downscale climate forcings from the regional to the catchment scale were either based on the weather generator described by Jones and Salmon (1995) and Watts et al. (2004) or the 'delta change' approach described by Hay et al. (2000). The weather generator was used at three sites (Esthwaite, UK; Flesk, IE; and Tarvastu, EE) and the delta change method at two sites (Arbogaån, SE and Mustajoki, FI). Multiple sequences of weather time series were created either directly from the weather generator or by re-sampling the monthly weather records and randomly recombining them. This allowed us to explore the uncertainty associated not only with the choice of model and scenario, but also in downscaling to the catchment scale.

For the CLIME sites in general, the projected DIN export closely follows the pattern set by streamflow. However, because streamflow is made up of surface runoff and baseflow and these flow paths are associated with different DIN concentrations in model input files, the results also reflect the seasonal variations in the proportion of these two streamflow components. For the purpose of comparison, loading estimates are normalised by area and expressed as fluxes in the following sections.

11.3.1 Model Projections for the Western Catchments (IE, UK)

All the climate change scenarios imply that Western Europe will become warmer, and in some cases wetter. Drier conditions were projected for late summer by all model combinations. On an annual basis, the increase in air temperature was greatest for the scenarios associated with the ECHAM4/OPYC3 model. At Lough Leane the average increase in the mean daily air temperature for the E A2 scenario was 3.6°C and the corresponding figure for Esthwaite Water was 5°. The differences between the sites are principally attributable to their locations. Lough Leane is near the coast in southwest Ireland in an area where winters are relatively mild. Esthwaite Water is situated farther north in the centre of the English Lake District where the winters tend to be much colder.

The results from the statistical downscaling of the RCM outputs with the weather generator reflect this geographical difference. Although the range of mean air temperature projections is narrower for the Flesk sub-catchment (ranging from 10.6°C for the control period to 14.2°C for the E A2 scenario), the average daily temperature is higher. The mean air temperatures for Esthwaite Water cover a greater range (9.3°C for the control period to 13.3°C for the E A2 scenario) but are more than a degree lower than those recorded in the Irish catchment.

Annual precipitation is adjusted in the model simulations to maintain a daily water balance. This value was higher in the Esthwaite catchment for the control period than in the Flesk sub-catchment (a median of 1898 mm as compared to 1551 mm) and is consistent with historical differences noted between the sites. At both locations, precipitation changes were greatest for the ECHAM scenarios, particularly for the E B2 simulations, with increases of 13 and 32% for the Irish and UK sites, respectively. The scenarios based on the Hadley model (H A2, H B2, Had A2, Had B2) show minimal changes (less than ±5%) on an annual basis for the Flesk catchment. However, there was a shift in the distribution of precipitation for the region. The marine climate of Ireland and the UK (Köppen classification Cfb) ensures that the winters will continue to be moist and mild, but the difference between wet winters and dry summers is more pronounced in the future climate. As a consequence, streamflow generally increased in winter and decreased in summer. For the Lough Leane catchment, mean monthly streamflow for all future

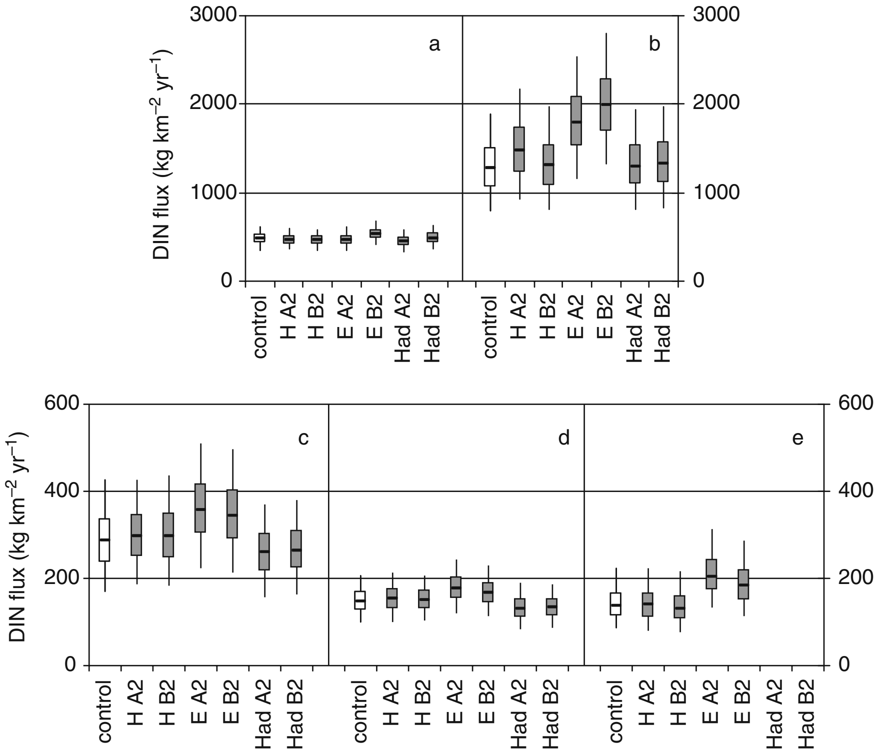

Fig. 11.2 DIN annual flux. All values based on the arithmetic average of median annual fluxes for 100 realisations of a 30-year time series for Western sites: **a**. Flesk (IE), **b**. Esthwaite Water and Northern sites: **c**. Mustajoki (FI), **d**. Arbogaån (SE), **e**. Tarvastu (EE)

Table 11.5 Annual percentage change in median dissolved inorganic nitrogen fluxes

	Climate model/scenario					
Site	E A2	H A2	Had A2	E B2	H B2	Had B2
Esthwaite (UK)	+43	+17	+4	+57	+4	+6
Flesk (IE)	–4	–2	–7	+10	–4	+2
Mustajoki (FI)	+25	+3	−10	+20	+3	+10
Arbogaån (SE)	+19	+2	−12	+12	+2	−11
Tarvastu (EE)	+50	+2	–	+34	–4	–

scenarios was lower from April through December, but the proportion of runoff to baseflow was also greater. For the Esthwaite catchment, summer flows have always been relatively low, and further reductions are suggested by all future scenarios.

The median DIN fluxes for multiple realisations of our simulated 30-year periods are shown in Fig. 11.2 and the pattern of change summarized in Table 11.5. In Fig. 11.2, the median values for annual means represent 100 simulations of 30 years (indicated by the centre line in each box); the interquartile range is shown by the box, and the whiskers show the 5 and 95 percentiles. The 'control' simulation represents present-day conditions. The percent change values in Table 11.5 are based on the average values recorded in the 'control' simulations i.e. the historical period.

On an annual basis, the E B2 scenario generated the greatest increase for both the Flesk (10%) and Esthwaite (57%) catchments. The Flesk simulations with the Had B2 scenario also showed an increase in annual DIN export but reductions were indicated for the H B2 scenario and for all three A2 scenarios. The Esthwaite catchment had the highest annual DIN export rates of the five catchments (Fig. 11.2), and the averages for the control period are very similar to those reported elsewhere in the UK (Hooda et al., 2000; Johnes and Butterfield, 2002). In the case of Esthwaite, there was greater variability between scenarios for yearly mean DIN flux but increased export rates were indicated for all scenarios.

In the Esthwaite catchment, the seasonal pattern of DIN flux closely followed the pattern established for total streamflow, with the greatest increases in winter (up to 114%) and decreases in summer (Fig. 11.3). The most striking change in DIN export was a reduction during May, June, and July, when modelled evapotranspiration exceeded precipitation in all future scenarios resulting in further decreases in streamflow.

The most important seasonal differences were those simulated for the Flesk catchment in the autumn when the contribution of runoff to total streamflow increased while the contribution from baseflow remained low. The net effect for all six scenarios is a decline in DIN flux for the autumn months continuing into December for all except the E B2 scenario. The partitioning of flow components

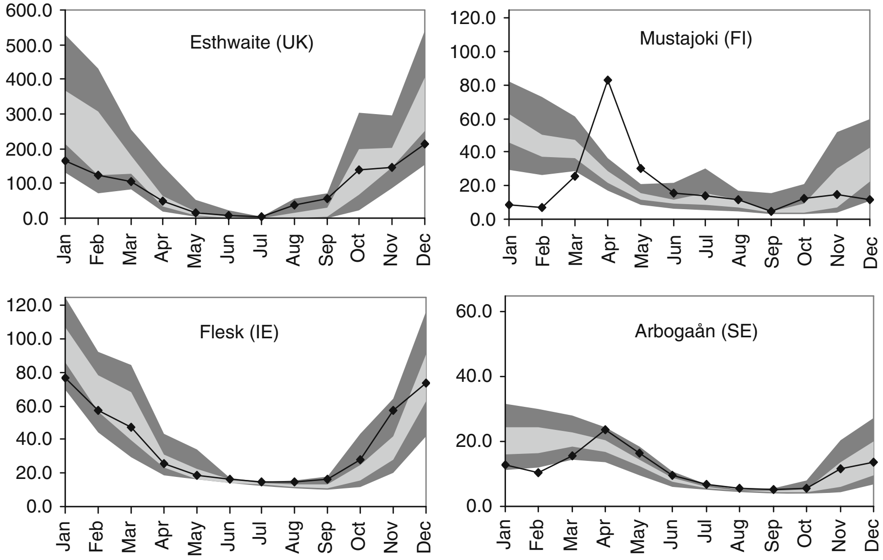

Fig. 11.3 Simulated DIN flux (median kg^2 km^{-2} mo^{-1}) for multiple realisations of a 30-year time series (*black line*: control period, 1960–1991; *light grey* envelope shows highest and lowest median values and *dark grey* envelope shows highest and lowest quartile values for six future (2071–2100) climate projections. Note scale changes on y-axis. Estonia is omitted because simulations were run for only 4 of 6 future scenarios

into surface runoff and baseflow is central to understanding the projected changes in autumn DIN fluxes at the Irish site. The simulations suggest a decline in DIN flux for the autumn and early winter period with a greater contribution from runoff and reduction in the contribution from baseflow.

11.3.2 Model Projections for Northern Catchments (FI, SE, EE)

The air temperature projections for the Northern catchments showed a marked increase in the winter with the most pronounced increases occurring in January for all scenarios examined. For the three Northern catchments, the annual increase in air temperature was higher for the A2 scenarios, with the E A2 scenario showing the greatest annual increase (5.0°C, 4.8°C, and 5.5°C for Finland, Sweden, and Estonia, respectively). Increased precipitation from January through March coupled with the reduction of water storage as snow resulted in a higher projected streamflow projected for this period in all scenarios. In most cases, the preceding winter months (November, December) also showed increases in streamflow.

The patterns of dissolved inorganic nitrogen (DIN) fluxes are best explained by the catchment water balance, described in detail in Chapter 3. Although the most extreme increases in air temperature are greater for the A2 scenarios irrespective of the climate model used, the net effect of precipitation and air temperature changes led to higher annual DIN flux estimates for the ECHAM model scenarios (E A2, E B2) (Table 11.5).

11.3.3 Comparisons with Other Model Projections

In the final stages of the project, we compared the GWLF results with the results of other modeling studies on the impact of climate change on the flux of dissolved inorganic nitrogen. In most cases, direct comparisons proved difficult, since these studies used different climate models, produced projections for different periods or were focused on geographic areas that were distant from our sites. A more systematic comparison was, however, possible at one CLIME site (Mustajoki) where the GWLF results were compared with some recent simulations with the process-based INCA-N model (Wade et al., 2002; Rankinen et al., 2004). This comparison showed that both models reproduced the correct annual DIN load. In contrast, the INCA simulations tended to overestimate the annual DIN load in years with high streamflow in late summer-late autumn (Bärlund et al., 2009). In an earlier study for Norway and Finland, also based on the INCA model but driven by the downscaled outputs from two Global Circulation Models, the INCA model projected substantial increases in winter streamflow (Kaste et al., 2004). Inclusion of biogeochemical cycling in this model showed increases in the N supply that was attributed to temperature and soil moisture related increases in net mineralisation. Here, changes in the mineralisation rate were balanced by an increase in N retention for the three catchments studied (Kaste et al., 2004).

The SWAT model (Soil and Water Assessment Tool; Arnold et al., 1999) was used to assess potential impacts of climate change on the supply of nutrients to the Yorkshire Ouse, an agricultural catchment in the UK (Bouraoui et al., 2002). Six climate scenarios were used, but the driving GCMs and the time slice (2050, 2080) differed from our study. The two most extreme scenarios used there also resulted in increased losses of total nitrogen during the winter but the explanation offered was an increase in the rate of mineralisation rather than an increase in streamflow. The SWAT model was also applied in Finland (Grizzetti et al., 2003) where a temporal analogue approach was used to evaluate climate impacts (Bouraoui et al., 2004). While their modelling approach was quite different from our own, this study also indicated that the increased winter runoff could be attributed to snowmelt and to the increases in temperature projected between January and April. A key finding of this study was that Nordic catchments can be highly sensitive to small variations in precipitation and temperature and an increase in total nitrogen loss from the catchment was primarily governed by the climatic factors that had the most pronounced effect on hydrological processes.

In Sweden, the impacts of climate change on nitrogen loading have been modelled for the Rönneå River of Lake Ringsjön in southern Sweden (Arheimer et al., 2005). Here, the SOILNDB (Johnsson et al., 2002) and HBV-N (Arheimer and

Brandt, 1998) models were used to simulate leaching and retention of nitrogen with six future climate scenarios. Four of the six scenarios used in this study were based on the same E A2/B2 and H A2/B2 scenarios used in CLIME. However, the method of transferring the climate signal to the water quality and quantity models (Andréasson et al., 2004) was different from that used in our study of the Galten sub-basin of Lake Mälaren. Arheimer et al. (2005) showed an increase in seasonality of streamflow and DIN loadings, which agrees with our results. In their model, the shift in N flux was mainly attributed to the increases in river discharge even when the scenarios were wetter and there was a decrease in the estimated N concentrations.

11.4 Discussion

In CLIME, we used the GWLF model to derive dissolved inorganic nitrogen projections for all the sites where there was sufficient supporting data. The structure of the model and the data requirements were discussed in a series of structured workshops to secure a consistent approach. The individual simulations were performed by staff in the partner countries to make sure that the best available information was used for the original calibration. This de-centralised approach facilitated the consistent application of the model to sites located in different climatic zones with very different land uses, soils, and catchment sizes.

The results summarised here suggest that the projected changes in the climate will have a major effect on the flux of DIN in all the selected catchments but the suggested patterns vary from site to site. The main contrast is between the sites located in Northern and Western Europe. For the UK site, annual dissolved inorganic nitrogen fluxes increased for all scenarios but the seasonal pattern of these inputs were very similar to those found in the control simulations. For the Irish site, the projected annual DIN fluxes tended to be lower than those in the control, with the exception of the Had A2 and Had B2 scenarios. The seasonal pattern of DIN flux followed the trend for baseflow for the autumn and early winter in the Flesk (IE) catchment. At the Northern sites, the key factor regulating the supply of DIN was the projected change in snow cover (see Chapter 10). At present, the maximum DIN loading in this region occurs during the dormant season. In the control simulations, designed to represent current conditions, the peak loads occur in April for Finland and Sweden and in March for Estonia. For example, in southern Finland, 75–90% of the DIN load is typically transported in the autumn and winter months, and early spring. In our simulation of future conditions this peak appeared one to three months earlier than that currently observed. Here, earlier snowmelt and a smaller proportion of winter precipitation falling as snow are also projected to shift peak streamflow from early spring to mid-winter.

In CLIME, we took advantage of the ‘dynamic hydrology’ feature of GWLF to run the very large number of simulations required to characterize the uncertainty associated with the DIN projections in the five, very different, catchments. The catchment-specific DIN concentrations are based on contemporary data and the hydrological components changed in a systematic way to reflect the projected range of the future values. This simplified approach deliberately avoids the problems posed by future changes in biogeochemical cycling or the effects associated with climate-related changes in land use. The results demonstrate that the dynamic changes in hydrology are well represented by the model and can be used with some confidence to explore the supply-related effects of the changing climate. Comparisons with other models suggest that GWLF is particularly good at simulating the projected changes in timing of nutrient delivery and the shifts that occur in the export of DIN. In the Western region, there was no major change in the timing of run-off but the projected increase in the rainfall meant that DIN loads increased in winter and early spring. In the Northern region, there was a pronounced shift in the timing of the snowmelt and runoff which greatly increased the DIN loads projected for the coldest months in the year. In Finland and Sweden the present April snowmelt peak disappeared altogether and was replaced by a more sustained increase in the DIN load projected for the winter. Although the Estonian site did not have a pronounced snowmelt peak, winter DIN loads also increased, due to the projected increases in winter precipitation.

Loss of snow cover may also result in increases in DIN flux following the freezing of soil, as low soil temperatures accompanied by changes in soil moisture have been shown to lead to increases in net mineralisation and nitrification rates and increased nitrate leaching losses (Groffman et al., 1999). While these effects are not represented in our modeling approach, they could augment the increases associated with the increased winter runoff.

While some studies link changes in DIN flux to accumulation in soils and biomass in dry years and flushing of the stored N pool in wet years (Donner et al., 2002; Andersson and Arheimer, 2001), others find no such relationship (Lucey and Goolsby, 1993; David et al., 1997). Howarth et al. (2006) suggest that nitrogen sinks are smaller in wetter catchments due to shorter water residence times in wetland and near-stream areas. This may, in turn, result in lower denitrification rates and lower retention. Increases in winter precipitation in both Western and Northern regions coupled with lower retention and denitrification could conceivably further enhance DIN export if the concentrations significantly exceed the averages used in our control simulations.

Changes in streamflow and nutrient inputs associated with future warmer world scenarios may also affect ecosystem function in the streams, rivers and lakes by modifying metabolic processes. The projected changes in flow regime and the timing of nutrient transport could also have a substantial effect on the growth rates and phenology of many aquatic species. A potential loss of hydrologic connectivity in the summer months is also anticipated if differences between evapotranspiration and precipitation increase to the extent suggested by the current climate scenarios.

Management of animal wastes and fertiliser application, as well as conservation tillage and changes in land use are among the available strategies adopted to minimise nitrogen pollution. River basin management through consideration of the timing and cycling of biogeochemical transformations (Schlesinger et al., 2006) could offset anticipated increases in nitrogen flux predicted by climate future scenarios.

The effects of land use change and other management strategies were not covered in CLIME. However, the modelling approach described here could be expanded to include such changes. For the larger marine basins draining into the Baltic Sea, a derivative of the GWLF model (CSIM, Mörth et al., 2007) has been used for such an assessment and incorporated into a decision support system (Wulff et al., 2007). In their management scenarios, the most pronounced reductions in nitrogen loading were achieved by converting agricultural land to forests on a Baltic-wide scale. Their modelling approach allowed them to explore the trade offs between different management scenarios for both P and N. Similar to the finding of Arheimer et al. (2005), multiple strategies were necessary to achieve reductions in nutrient loadings, including expansion of wetlands for nitrogen reduction.

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Chapter 12
Impacts of Climate on the Flux of Dissolved Organic Carbon from Catchments

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12.1 Introduction

Recent increases in dissolved organic carbon (DOC) concentrations in surface waters across both Europe and North America have focused attention on the factors controlling the export of DOC compounds from catchments. Waters containing high concentrations of DOC generally have a characteristic brown colour and are associated with the presence of highly organic soils. Catchments dominated by these soils typically export between 10 and 300 kg DOC ha^{-1} year^{-1} (Billett et al., 2004; Laudon et al., 2004; Jonsson et al., 2006). A portion of this DOC is mineralised in streams and lakes to CO_2, while the remainder is transported to the sea (Jonsson et al., 2006). Organic matter accumulates in soils when decomposition rates are restricted either by low temperatures or water-logged conditions. In Europe organic soils are found mainly in colder, wetter regions in the west and north (Montanerella et al., 2006) (Fig. 12.1). Peat soils are those with an organic content of greater than 25% (Montanerella et al., 2006). These highly organic soils represent a significant global carbon store (Gorham, 1991; Davidson and Janssens, 2006).

The regional nature of the recent increases in DOC concentrations suggests that large-scale drivers are involved. Both the production and transport of DOC are strongly influenced by climate and the observed increases have been linked, in part, to recent climate change. Suggested climatic drivers of the upward trends have included temperature (e.g. Freeman et al., 2001b; Evans et al., 2006), soil moisture impacts on decomposition processes (e.g. Worrall et al., 2006), solar radiation (e.g. Hudson et al., 2003), and variation in the timing and intensity of precipitation and snowmelt (e.g. Hongve et al., 2004; Erlandsson et al., 2008). These climatic impacts must also be viewed in the context of additional factors which influence DOC concentrations, in particular, recent decreases in anthropogenic acidification of surface waters associated with decreases in industrial emissions and subsequent sulphur deposition (Evans et al., 2006; de Wit et al., 2007; Monteith et al., 2007). Changes

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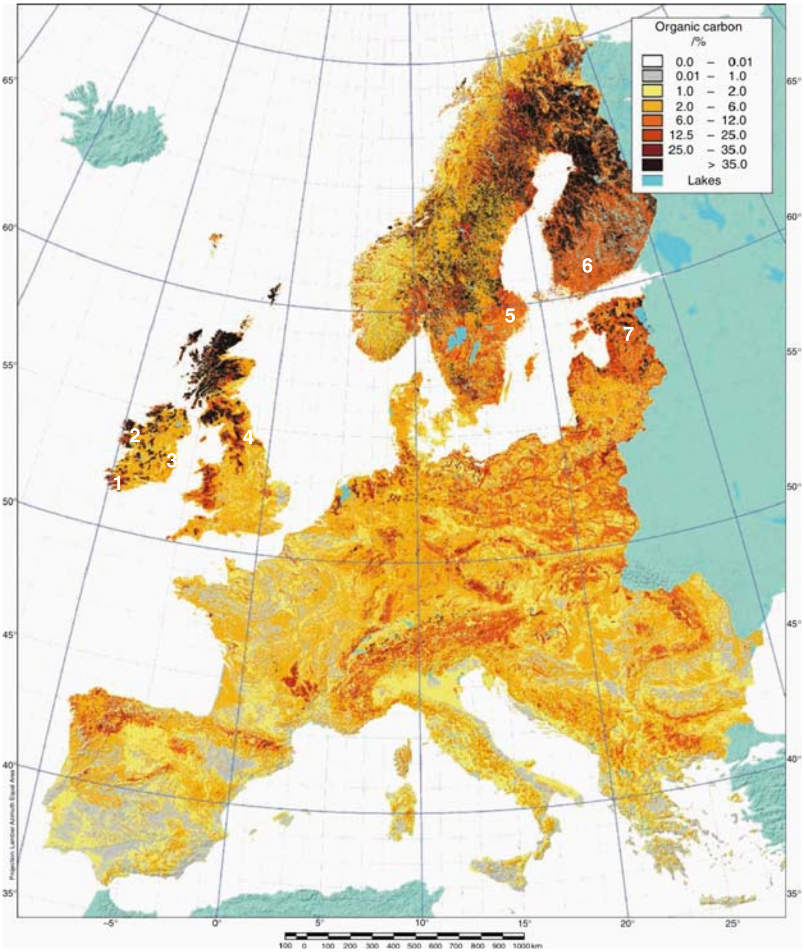

Fig. 12.1 Distribution of organic soils in Europe (from Montanerella et al., 2006). The annotations show the position of the CLIME sites used for the DOC studies described in Chapters 12 and 13 of this volume: 1 L. Leane, 2 L. Feeagh, 3 Poulaphuca Reservoir, 4 Moor House, 5 L. Mälaren, 6 L. Pääjärvi and L. Valkea-Kotinen, 7 L. Võrtsjärv. Map reproduced with kind permission of the European Commission Joint Research Centre (L. Montanterella)

in pH can influence the solubility of DOC compounds (Krug and Frink, 1983; Clark et al., 2005) and increases in DOC concentrations have been linked to decreases in sulphur deposition in the UK (Evans et al., 2006), Norway (de Wit et al., 2007), and Sweden (Erlandsson et al., 2008). An analysis of data from over 500 sites in North America and northern Europe has also reported that, for the period between 1990 and 2004, DOC concentrations increased in proportion to the rates at which atmospherically deposited anthropogenic sulphur and sea salt declined (Monteith et al., 2007). However, given the recent stabilisation in the levels of sulphur deposition (Skjelkvåle et al., 2005), and the potential sensitivity of DOC export to projected climate change (Sobek et al., 2007), climatic factors are likely to play a greater role in future variability and trends.

High DOC concentrations have implications for both water treatment and for the ecology of surface waters. DOC must be removed from drinking water because of health concerns related to the formation of trihalomethanes (THMs), carcinogenic compounds that are produced when water with a high DOC concentration is disinfected using chlorine (WHO, 2005). During the 1990s many water treatment plants in Nordic countries began to report an increased difficulty in treating highly coloured water (Löfgren et al., 2003; NORDTEST, 2003). Of particular concern has been the continued investment required to deal with the problem and indications that the quality as well as the quantity of organic matter appears to be changing (NORDTEST, 2003). Similar problems have been noted in the UK (Scott et al., 2001; Sharp et al., 2006). Changes in DOC concentration and water colour also have physical, chemical, and biological implications for lake ecosystems (Jones, 1998) while the export of DOC from catchments also represents a transfer of carbon from long-term terrestrial stores to more labile forms that can further contribute to atmospheric concentrations of CO_2 and, therefore, potentially contribute to global warming. In this chapter the influence of climatic factors on short-term variability and long-term trends in the export of DOC is described using examples from CLIME catchments and from other published studies. A related chapter (Chapter 13 of this volume) describes the model simulations that were used to quantify the future flux of DOC at CLIME sites given the climate change projections summarised in Chapter 2.

12.2 The Influence of Climatic Factors on DOC Production and Transport

12.2.1 Temperature

In general, the rate at which soil organic matter decomposes increases in response to a rise in temperature (Chapman and Thurlow, 1998; Davidson and Janssens, 2006). However, differences in the quality and availability of substrate, fluctuations in microbial populations and constraints on the access of microbial enzymes to substrate molecules contribute to a wide variability in reported temperature sensitivities (Kirschbaum, 1995; Chapman and Thurlow, 1998; Davidson and Janssens, 2006). Aerobic decomposition, which is restricted in some high organic soils to surface layers during periods of reduced soil moisture, is generally more responsive to temperature than anaerobic decomposition (e.g. Hogg et al., 1992; Weider and Yavitt, 1994; Davidson and Janssens, 2006). The relative response of decomposition rates can also be greater at lower than at higher temperature ranges (Kirschbaum, 1995; Chapman and Thurlow, 1998), while rates may also decline after a temperature optimum has been reached (e.g. Fenner et al., 2005). This temperature optimum has been noted to coincide with highest soil temperatures at different times of the year, suggesting some acclimation to ambient temperatures by the microbial population (Fenner et al., 2005). Similarly, acclimation to higher temperatures by soil microbes was suggested to account for observed reductions in decomposition rates during long-term studies in forested ecosystems, after initial increases in response to an experimental rise in temperature (Jarvis and Linder, 2000). However, recent results have indicated that the depletion of the more labile carbon stores in the soil was the cause of these reductions (Eliasson et al., 2005; Hartley et al., 2007).

The complex combination of factors influencing the rate of temperature response leads to a wide variation in reported temperature sensitivities, even in samples from similar locations (Kirschbaum, 1995; Chapman and Thurlow, 1998; Byrne et al., 2001; Davidson and Janssens, 2006). The rate at which soil organic matter decomposes is commonly assessed through measurement of the gaseous by-products of aerobic and anaerobic decomposition, which are CO_2 and CH_4 respectively. The impact of temperature on rates is often conveniently expressed as a Q_{10} value, the factor by which decomposition increases for a 10°C rise in temperature. Chapman and Thurlow (1998) reported Q_{10} values ranging from 2 to 19, with a mean value of 4.6 for decomposition in peat soils from sites across Scotland, (in the UK). The wide variation in the response to temperature was attributed, in part, to variation in soil moisture content, leading to differences in the contribution of aerobic and anaerobic decomposition. Landuse changes, such as afforestation and drainage, can also change the balance between anaerobic and aerobic processes and affect the apparent temperature response (e.g. Byrne et al., 2001). Decomposition measured as CO_2 flux from forested peat in the west of Ireland was 1.8–2.5 times greater than from virgin peat at the same location, with Q_{10} values of 3.0 and 2.6 for the two land uses respectively (Byrne et al., 2001) (Fig. 12.2). This difference in rates was linked to the impact of drainage at the forested site.

In contrast to the production of gaseous by-products, the concentration of DOC in soils and in streamwaters may not always show an immediate response to a

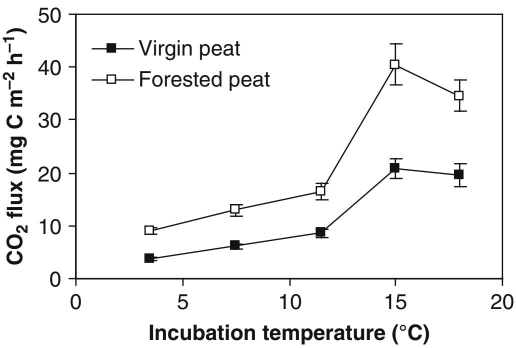

Fig. 12.2 The effect of an increase in incubation temperature on decomposition (CO_2 flux as mg C m^{-2} h^{-1}) measured in virgin and forested peat soils in western Ireland (after Byrne et al., 2001)

rise in temperature (Clark et al., 2005; Fröberg et al., 2006). A 4-week lag in the temperature response of soil DOC concentration was observed in a ten-year study in the UK, implying lags in either the activity of soil biota, population size of soil biota or the kinetics of DOC release (Clark et al., 2005). Despite this lag, there was a strong seasonal coupling between temperature and DOC, with temperature accounting for 58% of the variation in soil solution DOC concentrations.

12.2.2 Soil Moisture Levels

Decomposition rates are also highly responsive to changes in soil moisture, particularly in soils that are generally waterlogged. During dry periods the water table falls, allowing oxygen into anoxic layers and leading to an increase in aerobic decomposition. The concentration of DOC compounds in soil pore water may actually decrease during the drought itself and in the immediate post-drought period (Jensen et al., 2003; Freeman et al., 2004; Clark et al., 2005). Drought-induced acidification, mediated through the oxidation of organic sulphur to sulphate, and related decreases in the solubility of DOC, are thought to contribute to these lower concentrations (Clark et al., 2005). When the soil is then re-wetted following a prolonged dry period, increases in DOC concentrations in soil pore-waters have been reported in many studies (e.g. Fenner et al., 2001; Chow et al., 2006). The impact of low soil moisture levels on the production of DOC compounds has also been shown to be correlated with the extent of the time period of the preceding drought (Mitchell and McDonald, 1992). Rates, however, generally decrease in highly organic peat soils when the peat dries to the point where it becomes hydrophobic (Mitchell and McDonald, 1992).

Increases in surface water DOC concentrations have also been reported following drought (Naden and McDonald, 1989; Vogt and Muniz, 1997; Tipping et al., 1999; Fenner et al., 2001; Worrall et al., 2004). The positive effect of low moisture levels on stream water DOC concentrations can be prolonged, resulting in a step-like change which may persist for months or years (Naden and McDonald, 1989; Worrall and Burt, 2004; Worrall et al., 2004). These step changes may also be superimposed upon longer term decadal increases in DOC concentrations. Typical post-drought periods of elevated DOC concentrations in the Moor House catchment, a CLIME site in the UK, were 3–5 years (Worrall et al., 2004). Drought conditions may also impact cumulatively over several years (Naden and McDonald, 1989).

The exact nature of the positive effect of reduced soil moisture levels on decomposition rates is still not fully understood (Freeman et al., 2001b; Clark et al., 2005; Worrall et al., 2006). Freeman et al. (2001b) proposed an 'enzymic latch mechanism' to explain the effect. They showed that anoxic conditions in peat soils inhibit the enzyme phenol oxidase leading to an accumulation of phenolic compounds. These, in turn, inhibit the activity of the hydrolytic enzymes responsible for decomposition. The increase in oxygen in the surface peat layers results in a depletion of the stored phenolic compounds and allows the activity of hydrolytic enzymes to 'switch on'. Freeman et al. (2001b) suggested that, after a drought, the activity of the hydrolytic enzymes would continue until the concentration of phenolic compounds built up again and switched off enzyme activity. It has been proposed that this enzymic latch mechanism is the cause of step changes in DOC concentrations that have been reported following droughts in some catchments (Worrall et al., 2004). To date, reports of these prolonged upward shifts in DOC concentration have only come from peat catchments in the UK (Watts et al., 2001; Worrall et al., 2004). It is possible that the microbial community in drier climates reacts differently to decreased soil moisture (Jensen et al., 2003). A comparison of the effect of drought on DOC production from two heathland sites in Denmark and Wales, with annual precipitation of 750 and 1,700 mm respectively, reported decreased microbial activity and CO_2 emissions at the drier Danish site but increased activity at the wetter Welsh site (Jensen et al., 2003).

In order to explore the occurrence of step changes in DOC concentrations following summer drought in both Northern and Western Europe, time series from CLIME sites in Ireland, Sweden, Finland and Estonia (Table 12.1) were examined. Firstly, any step changes in DOC concentration were identified in the time series. The pattern of these step changes was then related to the occurrence of drought conditions at the selected sites. At three of these sites (the R. Fyrisån inflow to Lake Mälaren in Sweden, the Õhne sub-catchment of Lake Võrtsjärv in Estonia and Poulaphuca Reservoir in Ireland) the DOC data consisted of measurements of either water colour or potassium permanganate consumption (COD_{Mn}), both proxies for carbon content. At the fourth catchment (the Mustajoki inflow to Lake Pääjärvi in Finland) the only data available was a shorter time series of total organic carbon (TOC) measurements.

For each time series, a cumulative deviations plot was constructed (Buishand, 1982). In this test the cumulative deviations from the overall mean are calculated and rescaled using the overall standard deviation. Step changes in a cumulative plot of these values are indicated by a sharp change in direction and the statistical significance of the change point (i.e. the point with the maximum cumulative score)

Table 12.1 The characteristics of the sites used to quantify long-term change in the concentration of DOC (locations shown in Fig. 12.1). Air temperature and precipitation averages are for the period for which DOC data were available at each site

Site	Time period	Catchment km^2	Mean annual air temperature °C	Annual precipitation mm	Summer JJA precipitation mm
Fyrisån (Sweden)	1965–2002	2,005	5.9	557	177
Õhne (Estonia)	1980–2004	575	5.6	656	244
Poulaphuca (Ireland)	1977–2002	303	9.5	981	211
Mustajoki (Finland)	1993–2004	77	3.6	534	222

is assessed (Buishand, 1982). Since this test identifies only the largest step change in a time series, a further test for data sets in which multiple change points occur was applied (Lanzante, 1996). The standard precipitation index (SPI), a simple and effective tool for describing drought (Lloyd-Hughes and Saunders, 2002), was calculated for the months June, July and August as a measure of summer drought conditions using precipitation records from meteorological stations at, or close to, each site. The SPI is based on the cumulative probability of a given rainfall event occurring at a station and allows assessment of the rarity of a drought during the period of record. All drought events were identified at each of the four sites.

Significant step changes in DOC concentration were identified at both the Swedish and Irish sites. These were most notable at the Swedish site where upward shifts in concentration occurred in 1977 and in 1996 (Fig. 12.3a, c). Five summer droughts were identified at this site and both of the significant step changes in DOC concentration were preceded by one or more of these dry summers (Fig. 12.3c). The largest step change occurred in the autumn of 1976 and followed two sequential years with summer droughts. A second significant step change occurred in 1996, also following a summer drought. However, no change points were apparent after two other dry summers featured in the record. A significant step change was also identified at the Irish site, Poulaphuca Reservoir, in the autumn of 1984 and followed droughts in three out of the four preceding summers (Fig. 12.3b, d). A second but less significant change point was apparent in 1996, following the extremely dry summer of 1995. These step changes were most notable in the reservoir in the summer months, when lower concentrations are usually recorded.

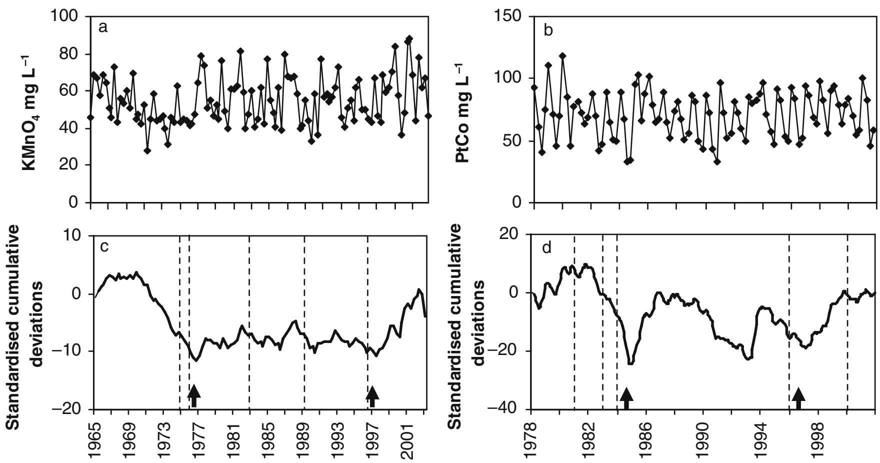

Fig. 12.3 Time series (**a** and **b**) and cumulative deviation plot (**c** and **d**) of permanganate oxygen consumption ($KMnO_4$mg L^{-1}) from the R. Fyrisån (Sweden) (**a** and **c**) and water colour (Pt Co mg L^{-1}) from Poulaphuca Reservoir (Ireland) (**b** and **d**). Significant step changes are indicated by *arrows*. Drought years as defined by the summer Standard Precipitation Index (SPI) (SPI value greater than −1) are indicated by the dashed lines

These results from Sweden are the only examples of step changes from a northern catchment to date and confirm that increases in DOC concentration do follow some, but not all, summer droughts in northern catchments. It is notable that the largest step changes at both the Swedish and Irish sites occurred following a sequence of dry summers. No significant step changes were found in the DOC data from the Estonian or Finnish sites despite the occurrence of summer droughts in several years at both sites. There were, however, no sequential summer drought years at these sites during the period examined. The changes at the Swedish and Irish sites also coincided with drought response years identified in several UK catchments (Watts et al., 2001; Worrall et al., 2004), highlighting the potential for large-scale drought to result in regionally coherent changes in DOC concentrations and the importance of taking account of such responses when analysing long-term trends.

12.2.3 Precipitation

The store of DOC compounds in peaty soils is flushed from the soil by rainfall or snowmelt, with the highest concentrations often coinciding with high flow levels (Andersson et al., 1991; Arvola et al., 2004; Worrall et al., 2002; Laudon et al., 2004). A decrease in concentrations at high flow, indicating a washout of soil carbon can, however, occur with successive events (Vogt and Muniz, 1997; Worrall et al. 2002). Concentrations can also show variation arising from the activation of different flow pathways. Worrall et al. (2002), working in the UK, identified three DOC fractions which were associated with different flowpaths. The first fraction was low in DOC and was related to rainwater which had little contact with the soil. The second also had low concentrations but originated from old groundwater which had largely been exhausted of DOC. A third fraction represented ‘new water’ but had high DOC levels supplied by the surface peats which had become a site of oxidation between events and, therefore, had a high supply of available carbon. Vogt and Muniz (1997) also reported high TOC concentrations in soil water from the upper peat layers overlying mineral soils in Norway. In contrast, a baseflow component that was associated with seepage from the mineral soil layer had lower DOC concentrations. In catchments where snow cover is a factor, the main export of DOC may be restricted to the period during or after the spring flood following snowmelt (Heikkinen, 1994; Laudon et al., 2004). Laudon et al. (2004) noted that a large percentage of spring meltwaters in a forested catchment in Sweden reached streams via subsurface pathways rich in TOC, whilst those in wetlands flowed over ice and frozen peat and had lower concentrations.

Investigations into the factors influencing the flux of DOC at CLIME sites have also highlighted the role of increased discharge in the wash-out of DOC (Weyhenmeyer et al., 2004b; Nõges et al., 2007; Arvola et al., 2004). A distinct increase in the loading of organic carbon into Lake Mälaren, Sweden, was noted in 2001 following exceptionally high rainfall in the autumn and winter of 2000

(Weyhenmeyer et al., 2004b). Water colour, measured as light absorption of filtered water, was up to 3.4 times higher in that autumn than at any time since records began in 1965. Examination of a 28-year record from the main inflows to Lake Võrtsjärv, Estonia, also found that both water colour and COD_{Mn} were positively related to discharge throughout the year, with significant correlations in the months of March and June and from August to November (Nõges et al., 2007). In addition, water colour and COD_{Mn} in spring were positively related to the North Atlantic Oscillation (NAO) in the previous winter, reflecting the occurrence of increased precipitation in high NAO index years. A positive relationship between the winter NAO and the TOC load in Finnish rivers has also been reported (Arvola et al., 2004).

To date, studies of the relationship between flow and DOC concentrations have mostly been based on data collected at monthly, weekly or, occasionally, daily time steps. In contrast, catchment responses to increases in discharge can be rapid, with peak streamflow often occurring within hours of a high rainfall event. Low frequency monitoring does not capture the initial impact of high flow rates on DOC export. However, in-situ instruments, which measure chromophoric dissolved organic matter (CDOM) fluorescence in stream water as a proxy for DOC concentration, were installed in catchments in both Ireland and Sweden, as part of the CLIME project. Measurements were logged every three minutes, averaged at hourly intervals and converted to standard quinine sulphate fluorescence units (QSU) (Fig. 12.4). The use of these high frequency measurements allows a unique assessment of the relationship between flow rates and concentrations at a sub-daily timestep, including the impact of snowmelt and storm events. The Irish site is on the Glenamong River, which flows into Lough Feeagh close to the Atlantic coast of Ireland (see Fig. 12.1). The area is characterised by high rainfall and snow is rare. Increases in fluorescence in the autumn of 2003 were associated with increases in flow, representing a flushing of CDOM stored in the soil matrix after the drier summer months (Fig. 12.4a). In the following February (julian day 31), an apparent dilution of baseflow concentrations occurred during high flow, indicative of lower carbon stores in the active flow pathways in the catchment (Fig. 12.4b). CDOM fluorescence decreased by over 20 QSU between two hourly measurements on the first day of this rainfall vent and then increased rapidly by a similar order of magnitude when baseflow was re-established. At the Swedish site, Hedströmmen, a river flowing into the western basin of Lake Mälaren (see Fig. 12.1), winter precipitation falls mainly as snow and there is little washout of soil carbon stores during winter months, although high frequency monitoring has shown that episodic increases in winter streamflow do occur. Here, increases in CDOM fluorescence were recorded in the period after the spring snowmelt (Fig. 12.4c) and during summer storms (Fig. 12.4d) (Moore, 2007). Peak river discharge after snowmelt occurred on March 21 (julian day 81), with CDOM concentrations increasing on the declining limb of the snowmelt hydrograph. Later in the same year (julian day 177–179), summer storms resulted in rapid increases in CDOM fluorescence. These increases were accompanied by high hour-to-hour variability in fluorescence.

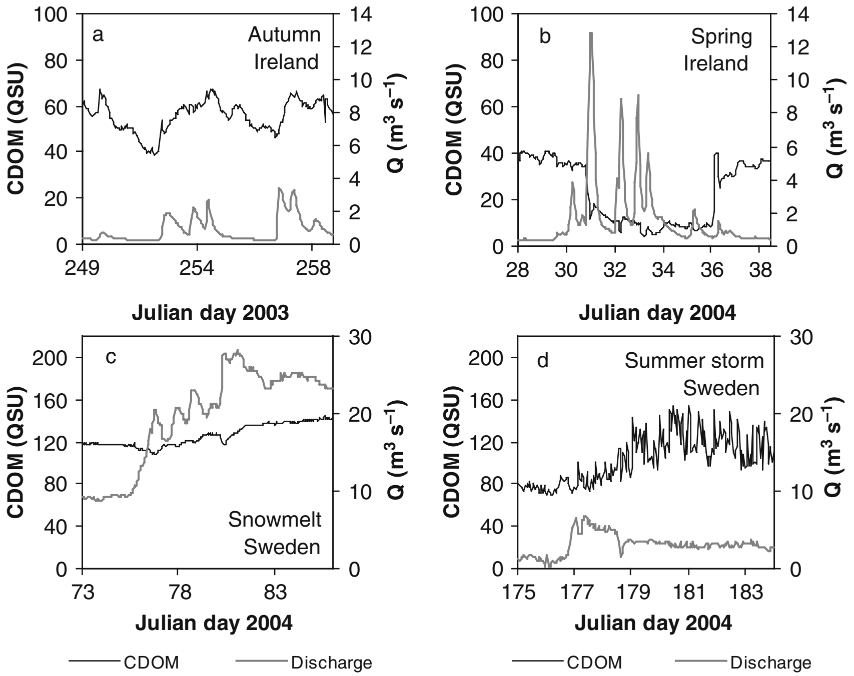

Fig. 12.4 Hourly CDOM fluorescence in quinine sulphate units (QSU) and river discharge (m^3 s^{-1}) from the Glenamong (Ireland) and Hedströmmen (Sweden) catchments. Note: QSU scales differ for the two sites

12.3 Additional Factors Influencing DOC Production and Transport

While climate has a strong influence on DOC production and transport, other factors also contribute to their seasonal and long-term variability. Seasonal changes in stream and lake concentrations are due, in part, to biotic and abiotic in-stream processing of DOC compounds (Schindler et al., 1997; Hongve, 1999; Köhler et al., 2002) while DOC derived from industrial activity and from agricultural and domestic waste also contribute to surface water concentrations in some areas (e.g. Apsite and Klavins, 1998). Land management can also impact on DOC production, particularly though changes in drainage. Drainage has been most extensive in Finland where approximately half of the total peatland area has been ditched (Kortelainen, 1999a). Drainage, together with the high water requirements of trees, lowers the water table, increasing aerobic decomposition in surface layers (Kortelainen, 1999a; Holden et al., 2004). Studies on the impact of drainage on DOC concentrations have been, however, contradictory with some studies reporting increases in concentration while others report decreases or no change at all (Holden et al., 2004). Deforestation, in contrast, can to lead to increases in DOC concentrations which may persist for several years (e.g. Neal and Hill, 1994; Cummins and Farrell, 2003). Variability in the concentration of DOC has also been related to changes in both natural and anthropogenic acidification in soils. Inverse relationships exist between both the pH and the ionic strength of soil porewaters and the release of DOC from soils (Thurman, 1985; Tipping and Hurley, 1988; Kalbitz et al., 2000). A decrease in soil pH following the oxidation of inorganic/organic sulphur stored in the peat was correlated with decreases in soil porewater DOC concentrations in a blanket peat catchment in the UK (Clark et al., 2005). However, while the soil was the main source of $SO_4{}^{2-}$ in very dry years, atmospheric deposition was the main source on other occasions. Anthropogenic acidification of soils and surface waters, associated with increases in sulphur emissions from industrial sources, increased during the 20th century and peaked in the 1970s. Rates have since reduced following the implementation of mitigation measures (Skjelkvåle et al., 2005).

12.4 Long-Term Trends in DOC Export

Increases in DOC and in proxies of DOC, such as water colour, have been reported from many catchments in Europe and North America (Freeman et al., 2001a; Skjelkvåle et al., 2001; Löfgren et al., 2003; Hongve et al., 2004; Evans et al., 2005; Worrall et al., 2005; Vuorenmaa et al., 2006). In the fifteen years up to 2004, DOC concentrations were shown to have increased significantly at 77% of 198 sites across the UK (Worrall et al., 2004), while concentrations doubled at all Acid Water Monitoring Network sites between 1988 and 2004 (Evans et al., 2005). Increases have also been widely reported from Swedish and Norwegian sites since the mid 1960s (Andersson et al., 1991; Löfgren et al., 2003; Hongve et al., 2004). Increases in water colour in Swedish lakes and rivers were initially noted for the period between 1965 and 1986 (Forsberg and Petersen, 1990; Andersson et al., 1991). A large-scale survey of 344 lakes in Norway, Finland and Sweden found increases in lake TOC concentrations in SE Sweden and southern Norway, but not in Finland (Löfgren et al., 2003) while Arvola et al. (2004) noted significant downward trends in TOC concentrations in nine of 16 rivers in Finland. More recently, TOC concentrations have been reported to have increased at ten small forested sites in Finland between 1987 and 2003 (Vuorenmaa et al., 2006). In North America, increases in DOC concentrations were observed in seven out of 48 lakes in the decade between 1992 and 2002 and in eight out of 17 lakes monitored from 1982 to 2002 in the Adirondack Region of New York (Driscoll et al., 2003). Increases in DOC were also noted in 17 out of 37 lakes in Quebec between 1985 and 1993 (Bouchard, 1997). In, contrast Schindler et al. (1997) reported a 15–25% decrease in DOC export to lakes in north-western Ontario between 1970 and 1990 which was attributed to decreased streamflow and to the effects of acidification.

At sites with longer records, the greatest increases have occurred in more recent decades (e.g. Löfgren et al., 2003; Hongve et al., 2004). An overall increase in water colour was observed at a Norwegian site between 1976 and 2002, but this included a decline from 1989 to 1991, and a steep increase between 1998 and the end of 2000 (Hongve et al., 2004). After 2000, water colour again declined. In addition, differences were observed between trends in water colour and DOC concentrations, indicating possible change in the quality of organic carbon over time. Declines were noted at three Swedish sites from the 1940s to the 1970s, with subsequent increases to 2002 (Löfgren et al., 2003). More recently, analysis of data from 28 Swedish rivers showed several periodic reversals in trend that were often synchronous across the region (Erlandsson et al., 2008). In all cases, the declines in concentration, centred around 1970, 1988 and 2004, coincided with periods of decreasing flow.

At the CLIME sites, increases in DOC concentrations have been reported for catchments in Sweden, Finland and the UK. A significant increase was found in absorbance data from three of the four sub-catchments of Lake Mälaren, Sweden, for the period 1965–2000 (Wallin and Weyhenmeyer, 2002; Tilja, 2003). While this increase was partially related to discharge, it could not be explained by variation in discharge alone (Tilja, 2003). A significant increasing trend has also been identified in the outflow stream from Lake Valkea-Kotinen, the secondary site used by CLIME in Finland (Vuorenmaa et al., 2006). In the UK it has been estimated that DOC concentrations increased by 0.1 mg C L^{-1} $year^{-1}$ between 1971 and 2002 at the Broken Scar Water Treatment Works, downstream from the Moor House catchment in the UK (Worrall and Burt, 2004). Long-term trends in the data from the four CLIME sites listed in Table 12.1, were assessed using the seasonal Kendall test (Hirsch et al., 1982). Significant upward trends were indicated at both the Irish and Swedish sites (Table 12.2). No significant trends were detected at the Estonian and Finnish sites. An increase of 0.017 mg C L^{-1} $year^{-1}$ was found at the Irish site, Poulaphuca Reservoir, for the period 1978–2001. The trend at the Swedish site for the period 1965–2002 was 0.17 mg $KMnO_4$ L^{-1} $year^{-1}$, representing an increase of 0.043 mg C L^{-1} $year^{-1}$. However, it is notable that if the period from 1972 to 2002 was used for the assessment a higher rate of change was indicated of 0.35 mg $KMnO_4$ L^{-1} $year^{-1}$, equivalent to 0.086 mg C L^{-1}. This time period ran from before the step change in 1976 described earlier to after the step change in 1996. Low concentrations prior to the 1976 step change and higher concentrations in the latter part of the record after the 1996 step change may have contributed to an apparent increase in the long-term trend at this site and illustrate that caution should be exercised when investigating trends in the presence of step changes.

Table 12.2 Trends in DOC at four CLIME sites assessed using the seasonal Kendall test. The trend is expressed in mg C L^{-1} $year^{-1}$ (NS = not significant)

Site	Period	Trend	Slope mg C L^{-1} $year^{-1}$	p
Fyrisån (Sweden)	1965–2002	positive	0.043	0.002
Õhne (Estonia)	1980–2004	NS		
Poulaphuca (Ireland)	1978–2001	positive	0.017	0.005
Mustajoki (Finland)	1993–2004	NS		

12.5 What Are the Drivers of the Observed Increases in DOC Concentrations?

There has been considerable debate regarding the main drivers of the recent increases in DOC concentrations, including the role of climate in these trends (e.g. Freeman et al., 2001b; Tranvik and Jansson, 2002; Freeman et al., 2004; Evans et al., 2005; Evans et al., 2006; Roulet and Moore, 2006; de Wit et al., 2007; Monteith et al., 2007). Freeman et al. (2001b) suggested that the increases could be due to the effect of temperature and soil moisture levels on the 'enzymic latch' mechanism described in Section 2.2. In reply, Tranvik and Jansson (2002) pointed out that the substantial increase in DOC concentrations in Swedish lakes and streams during the 1970s and 1980s had occurred despite a reduction in temperatures but was frequently linked to higher precipitation. Freeman et al. (2004) subsequently reported an increase in DOC export from peats, fens and riparian soils under experimental elevated CO_2 concentrations and suggested that the recent increase in atmospheric carbon dioxide levels was responsible for the observed higher DOC concentrations. However, as pointed out by Evans et al. (2005) atmospheric CO_2 increases have only been around 10% of the experimental increase used by Freeman et al. (2004), therefore only 1–6% of the actual increases in DOC concentrations could be explained by this mechanism.

Several authors have suggested that at least some of the upward trend may be due to decreases in anthropogenic acidification rates (Evans et al., 2006; de Wit et al., 2007; Monteith et al., 2007). A potential link between increased acidification and decreases in the solubility of DOC in soils was first suggested in the 1980s (Krug and Frink, 1983) and reductions in water colour in Norwegian streams in the period up to the 1960s were attributed to increased anthropogenic sulphur loading (Gjessing, 1970). Despite this, several large-scale catchment acidification experiments reported no obvious change in surface water DOC concentration with decreases in soil pH (e.g. Wright et al., 1993; Hessen et al., 1997). Analysis of more recent trends in DOC has again focused attention on possible links between changes in the rates of acid deposition and changes in the concentration of DOC in soils and surface waters. Recovery from acidification was among potential drivers reviewed by Evans et al. (2005), together with temperature, hydrological changes, drought-rewetting cycles, land-use change, in-lake and in-stream removal and nitrogen enrichment. The study concluded that deposition-related and climate-related factors both appeared to be significant and stated that, while it seems probable that recovery from acidification has contributed to DOC trends in those regions where sulphur deposition has decreased, comparable trends at unimpacted sites could only be explained by climatic factors. Evans et al. (2005) also noted that apparent step change increases following the dry summer of 1995 at a number of sites lent some support to the hypothesis that drought effects are involved. However, a further investigation concluded that declining sulphur deposition and changing sea-salt loading could account for the majority of the observed DOC trend in the UK (Evans et al., 2006). Reduced acid deposition has also been linked to increased DOC concentrations in lakes in

Norway (de Wit et al., 2007). Similarly, a recent large-scale study by Monteith et al. (2007) implied that most of the DOC increases reported in Europe and North America since 1990 can be attributed to decreases in anthropogenic acidification.

The role of both climatic factors and sulphate deposition has recently been investigated in an analysis of water colour trends in 79 reference lakes in Sweden between 1984 and 2004 (Weyhenmeyer, 2008). When these results were compared on a geographical basis (Fig. 12.5), there was a systematic north-south difference in the rate of change with a distinct discontinuity in the rates of change at latitudes above 60°N. This change point corresponds to the *limes norrlandicus*, the boundary between the more oceanic and more continental climatic regions of Sweden. The rate of change in colour was 3.5 times higher at southern latitudes than at more northern latitudes. North–south differences were also found for sulphate concentrations, which showed rates of change that were 2.7 times higher at lower latitudes. In contrast, the rate of change in sulphate wet deposition was only 1.5 times higher at lower latitudes, suggesting that factors other than deposition are contributing to the observed north-south difference in DOC to an almost equal extent. A similar north–south

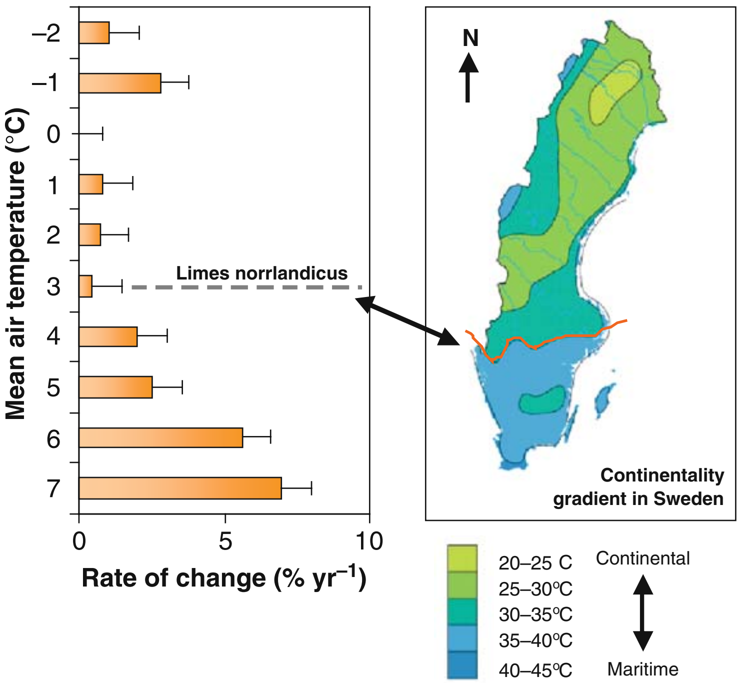

Fig. 12.5 Rate of change in water colour (estimated as light absorption at 420 nm of 0.45 μm filtered water in a 5-cm cuvette) in Swedish lakes along a N–S air temperature gradient. A distinct change was identified corresponding to the limes norrlandicus, the boundary between the more oceanic and more continental climatic regions of Sweden. Map reproduced with the kind permission of the National Atlas of Sweden

gradient and change-point have been observed for the timing of lake-ice break-up in Sweden, which shows significantly higher rates of change at lower than at higher latitudes (Weyhenmeyer et al., 2004a). The break-up of lake ice typically occurs when air temperatures exceed 0°C. The increase in temperatures above freezing point will also impact on catchment processes such as weathering rates, soil moisture levels and water flow characteristics, which may in turn affect DOC flux from the catchment. The relative importance of climate as a driver for changes in DOC concentrations in Swedish surface waters has also been highlighted by Erlandsson et al. (2008) and has implications for future management of water resources in the region, given that the impacts of climate change are projected to be more pronounced at higher latitudes.

12.6 Ecological Implications of High DOC Concentrations in Surface Waters

Increases in the flux of DOC from catchments can have a major effect on the ecology of downstream waters. Dark water absorbs more of the incident solar radiation, leading to an increased extinction of light and a shift towards the red part of the spectrum in many humic lakes (Eloranta, 1978; Jones and Arvola, 1984) (see Chapter 6 this volume). DOC compounds thus compete with phytoplankton for available light and can restrict photosynthetic production (Jones, 1992). The absorption of in-coming light by DOC compounds can also result in steeper thermal gradients, shallower thermoclines and increased stability, particularly in small, sheltered lakes. Primary production in these lakes is typically limited to the uppermost zone of the water column because of this steep stratification and may be little affected by any change in colour (Jones, 1992; Arvola et al., 1999). Many highly coloured lakes also exhibit net heterotrophy due to the central role of allochthonous DOC in fuelling the upper trophic levels via the bacterioplankton-protozoan link (e.g. Tranvik, 1992; Kankaala et al., 1996). However, in small and sheltered humic lakes the hypolimnion can also easily become anoxic during stratified periods as DOC is decomposed (Salonen et al., 1984).

High DOC concentrations can also impact on nutrient availability. In the boreal region, highly coloured lakes may stratify in the spring immediately after ice-melt, preventing the supply of hypolimnetic nutrients to the epilimnion (Salonen et al., 1984). Metal binding properties of humic substances and the interaction of humus-iron complexes with phosphate can further reduce the concentration of dissolved nutrients (Francko and Heath, 1983), but also act as nutrient reservoirs during periods of low availability (Francko and Heath, 1983; Jones, 1998; Vähätalo et al., 2003). Dissolved humic substances with high concentrations of organic acids contribute to the naturally low pH observed in humic waters (Kortelainen and Mannio, 1990; Lydersen, 1998). In contrast, they also markedly contribute to buffering capacity, which can mitigate the effects of acidification (Johannessen, 1980; Kortelainen, 1999b). Organisms living in humic waters can better withstand the low pH, because humic substances reduce the toxic effects of Al and other metals (Hörnström et al., 1984). Allochthonous DOC also has a central role in fuelling the upper trophic levels via the bacterioplankton-protozoan link in humic waters (e.g. Tranvik, 1992; Tulonen et al., 1992; Kankaala et al., 1996) explaining the net heterotrophy of brown-water lakes (Jansson et al., 2000) In-stream processes, such as microbial decomposition and photo-degradation, modify DOC quality and affect bioavailability in the recipient water body (Köhler et al., 2002).

12.7 Conclusion

Much of western and northern Europe is covered by soils with a high organic matter content. These soils represent a significant global store of carbon and the rate at which DOC is exported from these catchments has major implications for biogeochemical and biological processes in catchment soils, downstream waters, the treatment of potable water, and the global carbon cycle. While DOC production and transport can be influenced both directly and indirectly by factors, such as land use and acid deposition, climate plays a significant role in both short-term and long-term variability. The results presented in this chapter demonstrate that the impact of these climatic factors on DOC production and transport is complex and includes the combined effects of both temperature and precipitation on the decomposition, solubility and hydrological transport of these compounds. Although decomposition rates are responsive to changes in temperature, there is huge variability in reported temperature sensitivities. In addition, while increases in DOC concentrations have been shown to occur after summer droughts in northern as well as western catchments, this response is not apparent in all catchments nor does it occur after all drought years. The high resolution monitoring from Swedish and Irish catchments illustrate the variability on both seasonal and sub-daily timescales in DOC export and the complex interaction between the build-up of soil carbon stores, flushing rates, and the activation of hydrological pathways.

These climatic drivers also interact with anthropogenically mediated influences on DOC export and recent studies have highlighted the potential role of changes in sulphur deposition in observed upward trends. Given the potential complexity of the interaction between landscape, climatic and anthropogenic impacts, however, Roulet and Moore (2006) state that caution should be exercised in attributing the increases in DOC concentrations to any single factor. While changes in acid deposition may have played a role in the trends observed, the recent stabilisation in the levels of sulphur deposition (Skjelkvåle et al., 2005) would indicate that this factor will be of lesser importance in the coming decades. In contrast, as illustrated in this chapter, climatic factors play a major role in short-term variability and long-term trends in the export of DOC from catchments. The potential sensitivity of DOC export to climate change has been highlighted in a recent study of DOC concentrations in over 7500 lakes in six continents (Sobek et al., 2007). Given the projected changes in climate of Europe, described in Chapter 2 of this volume, the importance of climatic factors as drivers of future variability and trends can only increase. In addition, the export of DOC represents a major transfer of organic carbon from terrestrial stores to more active dissolved forms, and ultimately to the atmosphere as CO_2 (Gorham, 1991; Kirschbaum, 1995; Evans et al., 2005). A positive feedback mechanism between decomposition in organic soils and climate change has been suggested, with changes in climate driving increases in the export of DOC which in turn would increase atmospheric CO_2 concentrations and contribute to further global warming (Knorr et al., 2005; Davidson and Janssens, 2006).

Projected changes in the climate of Europe include increases in temperature in all regions, with implications for the rate at which soil organic carbon stores are decomposed. Significant decreases in summer rainfall are also projected for western catchments, further increasing the possibility of drought and subsequent impacts on the balance between the rates of anaerobic and aerobic decomposition, microbial enzyme activity and the solubility of DOC compounds in the soil. In addition, higher winter rainfall and projected changes in the timing and magnitude of soil freezing and snowmelt in northern catchment have the potential to lead to major changes in the seasonal pattern of DOC export in that region. The possible impact of these changes on DOC export at CLIME sites is explored in Chapter 13 of this volume in which climate-induced responses in DOC production and transport are modelled using the climate scenarios described in Chapter 2.

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Chapter 13
Modelling the Impacts of Climate Change on Dissolved Organic Carbon

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13.1 Introduction

Dissolved organic carbon (DOC) from peat soils has implications both for the ecology of receiving waters and for the quality and treatment costs of water used for human consumption. Fluxes of DOC from peat soils are also relevant in the context of the global carbon cycle. Chapter 12 in this volume has reviewed the evidence for the effects of different environmental factors on the decomposition of peat soils and the export of DOC, drawing on literature and long-term data acquired from a number of European sites. The conclusion from this and many other studies is that, although there may be other influences such as land management and recovery from acid deposition, climate factors are a major player in both the short-term variability and longer-term trends seen in measured DOC concentrations and fluxes. Given the importance of DOC and likely future changes in climate, it is timely and opportune to make use of our current understanding to project possible future DOC.

In this chapter we, therefore, focus on modelling climate-induced changes in DOC under the range of climate scenarios described in Chapter 2. We describe the modelling approach and its application under both current and future climate. Use of a number of catchments and climate scenarios affords an initial indication of the range of projected changes arising from different model parameters, greenhouse gas scenarios and climate forcing. The examples are chosen from both Western (Ireland and the UK) and Northern (Finland and Sweden) Europe. While indicative of future changes, the simulations necessarily reflect our choice of modelling approach and its ability both to represent current process understanding at the catchment scale and to simulate contemporary DOC concentrations and fluxes.

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13.2 Modelling Approach

In seeking to represent the potential effect of climate change on both concentrations and fluxes of DOC, there are many possible model structures that could be explored. Existing models fall into two groups. Models in the first group focus on DOC in layered soils (Taugbøl et al., 1994; Tipping, 1996; Neff and Asner, 2001; Michalzik et al., 2003). They include temperature-controlled organic matter decomposition, sorption reactions and a variety of hydrological formulations. DyDOC (Michalzik et al., 2003), the latest example of these models, considers three different humic fractions and three different soil horizons giving a large number of model parameters which either require measurement or calibration. None of the models are specifically formulated for peat soils. The second group of models are those which were originally developed to look at long-term peat accumulation (Clymo et al., 1998; Frolking et al., 2001). These models deal with the decomposition of plant material and the development of deep peats over centuries or millenia. They have tended to use the traditional description of a peat profile with an upper acrotelm which is seasonally drained and a lower catotelm which is permanently wet. They are focused primarily on the soil column rather than the catchment and do not explicitly simulate DOC.

Within the context of CLIME, the requirement is for a relatively simple, yet generally applicable, catchment-scale DOC model which is compatible with the formulation of the GWLF hydrology (described in Chapter 3). It should also include the climate influences on peat decomposition and DOC export discussed in Chapter 12. None of the existing models fulfils these requirements. The model developed within CLIME uses a simple two-phase representation of the production and washout of DOC as shown in Fig. 13.1. In terms of complexity, the model, therefore, lies between the simple mixing model approach, with specified concentrations for different land uses, taken in the nutrient modelling (Chapters 9 and 11), and the more detailed published DOC models which explicitly include multiple carbon pools and sorption-desorption reactions to describe chemical mobility. The model is focused at the catchment scale and, while not spatially explicit, it is anticipated that lateral, rather than vertical, flow processes play a dominant role in determining the export of DOC from the catchment. Washout is, therefore, modelled as a function of the store of DOC available for export and the water flux in the different hydrological pathways.

13.2.1 Production of Dissolved Organic Carbon

Dissolved organic carbon is produced largely through the microbial decomposition of organic material. In developing a simple model for peaty catchments on decadal time scales, we have assumed that the production of DOC is not limited by the amount of organic matter and so is considered independent of the total carbon store. This assumption is perhaps debatable in the context of forest soils – for example, on the basis of ^{14}C evidence, Schiff et al. (1997) show that most DOC is quite

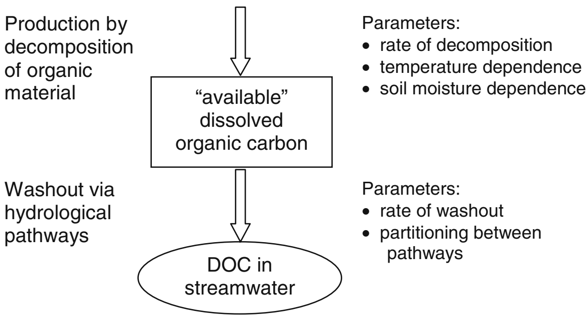

Fig. 13.1 Two-phase process of production and washout represented in the model

young (i.e. within the last 40 years) and Tipping et al. (2005) suggest that 30% DOC comes from the relatively small litter pool. However, by definition, peat forms through an excess accumulation of organic material suggesting that source limitation is unlikely. With regard to the amount of DOC produced, we know that both temperature and soil moisture can be important controls on the decomposition rate (see Chapter 12).

The dependence of decomposition, and hence DOC production, on temperature is widely accepted based on experimental evidence from both CO_2 (Chapman and Thurlow, 1996; Silvola et al., 1996; Chapman and Thurlow, 1998; Byrne et al., 2001; Dioumaeva et al., 2003) and DOC measurements (Christ and David, 1996; Moore and Dalva, 2001; Clark et al., 2006). It is often described in terms of the kinetics of enzymatic reactions using an Arrhenius equation (Taugbøl et al., 1994; Worrall et al., 2004) or Q_{10} relationship (Michalzik et al., 2003). Here, we use the formulation given in Equation (13.1). This has the advantage that the parameter used is the activation energy rather than the widely quoted, but temperature-dependent, Q_{10} value (Davidson and Janssens, 2006).

$$D_T = D_{T_{ref}} \exp\left(\frac{E_a}{R}\left[\frac{1}{T_{ref}} - \frac{1}{T}\right]\right) \tag{13.1}$$

where D_T is DOC production rate (gC/m^2/day) at soil temperature T, D_{Tref} is DOC production rate at a reference soil temperature (gC/m^2/day), E_a is activation energy (kJ/gC), R is the universal gas constant (6.928×10^{-4}kJ/K/gC), T is soil temperature (K), T_{ref} is reference soil temperature (K).

Dependence of decomposition on soil moisture is less well established than the dependence on temperature. Drying and wetting experiments on UK peat cores by Mitchell and McDonald (1992) showed total DOC production to be a function of soil moisture, with maximum DOC export from cores which had attained a soil moisture deficit of around 35% and much lower values from both wetter and drier cores. More recently, experimental work by Clark et al. (2006) showed that DOC production in peat cores increases with increasing water table depth. From collated CO_2 measurements across Finland, Silvola et al. (1996) also showed that, once values have been standardised for temperature, decomposition increases with increasing water table depth and then decreases with a further increase in water table depth. Christ and David (1996) similarly show increases in dissolved organic carbon with increasing moisture at very low moisture levels.

Formulating an equation to represent these findings requires both a consistent measure of soil moisture and compatibility with the hydrology model being applied. Water table depth is not generally transferable between sites or soil types so some form of volumetric measure must be used. The component of soil moisture which is represented in the GWLF hydrological model (see Chapter 3) is the unsaturated zone i.e. the soil moisture content between field capacity and wilting point. Following Päivänen (1973), the unsaturated zone has been assumed to represent about 50% of the total soil moisture within peat soils, with residual soil moisture providing another 30%. Given this, a suitable equation derived from the Mitchell and McDonald (1992) experiments is

$$D_S = a \left\{ \exp \left(-b \left[\left(\frac{(S_{\max} - S)}{1.6 S_{\max}} - 0.35 \right)^2 - 0.1225 \right] \right) \right\} \tag{13.2}$$

where D_S is DOC production rate (gC/m^2/day) at soil moisture S, a is anaerobic DOC production rate (gC/m^2/day), b is the rate of change in DOC production with soil moisture, S_{max} is the soil water capacity of the unsaturated zone (cm) and S is the soil water content of the unsaturated zone (cm). Assuming a relationship between soil moisture and water table depth (e.g. Scholtzhauer and Price, 1999), this equation may be shown to be consistent with the data of Clark (2005) and Silvola et al. (1996). It is fundamentally different to the linear dependence assumed in the INCA-C model (Futter et al., 2007).

The equation for soil moisture dependence now needs to be coupled with the temperature dependence. In the Mitchell and McDonald experiments, the cores were subjected to ambient air temperatures for different periods of time in order to induce drought conditions. Thus, temperature was only loosely controlled. If we assume that Equation (13.2) applies for a given temperature and that the effect of soil moisture is to limit the availability of substrate for microbial activity through the exposed surface area of pore space within the peat, then the soil moisture dependence should have a multiplicative effect on the temperature dependence. Indeed, this provides a first approximation to coupling the Arrhenius and Michaelis-Menten kinetics as suggested by Davidson and Janssens (2006). This formulation is also supported by observations that aerobic decomposition is more sensitive to changes in temperature than anaerobic decomposition (Hogg et al., 1992; Chow et al., 2006; Evans et al., 2006). In the model, DOC production is therefore given by

$$\begin{aligned} D_{T,S} = a_{T_{ref}} & \left\{ \exp \left(-b \left[\left(\frac{(S_{\max} - S)}{1.6 S_{\max}} - 0.35 \right)^2 - 0.1225 \right] \right) \right\} \\ & \exp \left(\frac{E_a}{R} \left[\frac{1}{T_{ref}} - \frac{1}{T} \right] \right) \end{aligned} \tag{13.3}$$

where $D_{T,S}$ is DOC production (gC/m^2/day) at soil temperature T and soil moisture S and a_{Tref} is anaerobic DOC production rate (gC/m^2/day) at the reference soil temperature.

Implementation of this equation within GWLF, requires an estimate of soil temperature. In the absence of data to calculate a full energy balance and in keeping with the nature of the hydrological model, soil temperature has been estimated by a 10-day moving average of the air temperature above zero. Decomposition is assumed to cease when soil temperature is zero. As soil warming in spring is generally observed to be much slower than the corresponding cooling effect in winter, the moving average is taken over an extended 30-day period for the two months after snowpack reduction below 2 cm. This gives good agreement with available soil temperature data.

13.2.2 Washout of Dissolved Organic Carbon

The production of DOC is treated in terms of a single carbon pool for the catchment as a whole. Export of DOC from this pool is assumed to occur through the interaction of the hydrology and the store of available DOC. In reality, this interaction will be mediated by chemical processes including sorption-desorption and mineralisation (Tipping and Woof, 1991). However, it is expected that these processes are less important in peat soils and, at a simple level, we, therefore, assume the DOC store to represent the net DOC available for leaching. At the catchment scale, the washout of DOC is then assumed to be adequately simulated by a series of first order rate equations representing each of the hydrological pathways. There are three hydrological pathways described within our application of the GWLF hydrology model: runoff which, for peaty areas, is interpreted as rapid near-surface flow; rapid subsurface flow which is fed from percolation when the unsaturated zone is full; and slow subsurface flow which is fed via deep percolation and maintains river baseflow during dry periods (see Chapter 3 for further details). For each of the three pathways, operating in parallel, the amount of DOC which is exported is essentially calculated as a function of the total available DOC store, the water flux in the pathway and a rate constant. The DOC concentrations in each pathway, therefore, vary predominantly with the available DOC; the DOC concentration at the catchment outlet varies not only with the available DOC but also with the mix of waters from different pathways and, in the case of runoff, from different land covers. In-stream carbon processing is not included.

Runoff, or rapid near-surface flow, is defined for each land cover. We therefore need to define which land covers, taken as a surrogate for soil type, will produce DOC, and then use the runoff for these land covers to drive the DOC export via this pathway. For the catchments being modelled (see Table 13.1), the CORINE land cover classes of unexploited and exploited peat bogs, inland marshes, coniferous forest, mixed forest, natural grasslands, moors and heaths, and transitional woodland-shrub are regarded as producing DOC. It is assumed that no DOC is

Table 13.1 The catchments used for the DOC modelling

Catchment	Map no.*	Area km^2	Elevation m a.s.l.	Dominant soil types	CORINE landcover (classes >15%)	DOC area %
Glenamong Ireland	2	18.2	10–600	Peat	70% peat bogs 26% forest	99.9
Upper Catchment Lough Leane Ireland	1	125	50–1000	Peat Peaty podzols	43% peat bogs 37% moors & heaths	85.6
Trout Beck UK	4	11.4	535–848	Peat Peaty podzols	68% peat bogs 25% moors & heaths	100
Mustajoki Finland	6	76.8	103–180	Peat Peaty podzols	67% coniferous forest 20% peat bogs	86.9
Hedströmmen Sweden	5	998	6–291	Iron podzols Brown forest soils lithosols	53% coniferous forest 19% woodland/shrub	80.6

* see Fig. 12.1 Chapter 12

exported from other land covers, although, in practice, small amounts of DOC are also leached from agricultural land. As runoff from DOC-producing land covers is generally higher than from other pervious land covers, due to their higher curve numbers (see Chapter 3), this has a non-linear effect which is not proportional to the area of the catchment occupied by DOC-producing land covers. In addition, there is some evidence to suggest that soil moisture may be a control on export in terms of the connectivity of pore spaces within the peat (Mitchell and McDonald, 1992). Dilution during periods of high rainfall (Worrall et al., 2002; Clark et al., 2008) or snowmelt (Laudon et al., 2004) are also reported in the literature. In order to approximate this, we assume that above a certain rate of runoff, R_0, there is no additional washout of DOC and, as a consequence, dilution occurs. Including these secondary controls, gives the equation:

$$W_{runoff} = k_{fast} C \ \min(R_{luDOC}, R_0) \frac{S}{S_{\max}} \tag{13.4}$$

where W_{runoff} is washout via runoff pathway (gC/m^2/day), k_{fast} is rate of washout (fraction/cm), C is available DOC within the soil (gC/m^2), R_{luDOC} is runoff from DOC-producing land covers (cm/day), R_0 is the threshold above which no further washout of DOC occurs, S/S_{max} is an adjustment for soil moisture which varies between 0 and 1. In the majority of cases, R_0 was set to R_{luDOC} as the data available did not support calibration of this parameter.

Within GWLF, the subsurface hydrology is only considered at the level of the whole catchment and, with the exception of the unsaturated zone, it is pathways rather than storages of water which are represented. In deep peats, this lumped approach may suffice and washout via subsurface flows is given by

$$W_{ssf} = k_{slow}\ C\ (P + P_{deep}) \tag{13.5}$$

where W_{ssf} is washout via subsurface flow (gC/m^2/day), k_{slow} is rate of washout (fraction/cm), C is available DOC within the soil (gC/m^2), P is percolation (cm/day) and P_{deep} is deep percolation (cm/day). However, we have also applied the model to northern catchments, where forest is often found on soils with an upper peat layer and a lower layer of mineral soil. The concentrations of DOC within the pore waters of the two layers can differ substantially. For example, Starr and Ukonmaanaho (2004) quote DOC concentrations of 7.2–36.0 mg/l at 15 cm and 4.1–21.2 mg/l at 35 cm for undisturbed boreal forest ecosystems. It is also known that DOC can be adsorbed onto mineral matter in the lower soil horizon and that this can lead to different climate responses for different soil types (Tipping et al., 1999). As a simple approximation for this effect and assuming that the DOC production given by Equation (13.3) is the net amount available for leaching, an adjustment factor to represent the area of the catchment occupied by soils with a lower mineral horizon was applied to the DOC washout via deep percolation:

$$W_{slowssf} = k_{slow}\ C\ P_{deep}\ \left(\frac{MA_{\min} + A_{peat}}{A_{luDOC}}\right) \tag{13.6}$$

where $W_{slowssf}$ is washout via slow subsurface flow (gC/m^2/day), M is an adjustment for subsurface mineral soils, A_{luDOC} is the area of DOC-producing land covers (km^2) with A_{min} being the area of soils having a lower mineral horizon (km^2) and A_{peat} being the area of deep peats (km^2). This allows the model to simulate very low DOC concentrations during times when streamflow is solely due to slow subsurface flow.

The available DOC within the soil is updated in each time step on the basis of the carbon mass balance between production and washout. The final DOC concentrations in streamflow are calculated from all the contributions from subsurface and near-surface flows, including the non-DOC producing land covers, using a simple mixing model i.e. total DOC flux divided by total water flux.

13.3 Modelling Dissolved Organic Carbon Under Current Climate

The coupled GWLF-DOC model is driven by daily temperature and precipitation data. The strategy for modelling the impact of climate change is first to calibrate the model for current hydrology and DOC using available data for streamflow and DOC concentration. Chapter 3 discussed the application of the GWLF hydrology model. Here, we focus on the DOC model. Details of the catchments considered are

Table 13.2 The DOC data available for the modelled catchments

Catchment	Data available	No. values	No. years	Frequency
Glenamong Ireland	DOC[†] Aug 2003–Mar 2005	242	2.5	Daily with gaps
Upper Catchment Lough Leane Ireland	DOC[†] 1999–2003	152	5	Weekly with gaps
Trout Beck UK	DOC 1995–2000	305	6	Weekly
Mustajoki Finland	TOC[‡] 1994–2002	325	9	Weekly with gaps
Hedströmmen Sweden	DOC* Sep 2003–Dec 2005	785	2.2	Daily with very few gaps

[†]DOC estimated from a regression on water colour ($R^2 = 0.98$; $n = 50$)
[‡]DOC approximated by total organic carbon (TOC) measurements
*DOC derived from hourly coloured dissolved organic matter (CDOM) corrected for temperature-mediated quenching of the fluorescence signal (Moore et al., pers. comm.)

given in Table 13.1. Their location, according to number, can be found in Chapter 12 (Fig. 12.1). Calibration of the hydrology (see Chapter 3) gave a reasonable to good fit in all the catchments with Nash-Sutcliffe efficiency values (Nash and Sutcliffe, 1970) between 0.57 and 0.78 and a realistic identification of the relative contribution of flow via near-surface and subsurface pathways. Information on the available DOC data is given in Table 13.2 .

In calibrating the DOC component of the model, a number of different objective functions and staged procedures were investigated. The method which was adopted was minimisation of squared deviations from the time series of DOC concentrations, rather than any derived set of statistics (e.g. percentiles) or calculated fluxes. The model was calibrated within the Vensim package using a modified Powell search of the parameter space between defined limits either for all parameters simultaneously or in a sequential optimisation. For the sequential optimisation, the observed DOC time series was used in conjunction with the baseflow-separated streamflow (Arnold and Allen, 1999) to give DOC concentrations in baseflow, very low baseflow (< 10–20% streamflow) and, using a mixing model and linear interpolation of baseflow concentrations, runoff. An initial estimate of the washout ratio (k_{fast}/k_{slow}) for the catchment was provided by the mean ratio of the derived concentrations in runoff to those in baseflow. An estimate of the mineral soil adjustment M was taken from the literature as the ratio of the DOC concentration in upper peat layers to that in lower mineral soil layers. These initial values were used in the first step of the optimisation in which a_{279}, b, E_a and k_{slow} were calibrated against the observed time series of DOC concentrations in streamflow. Revised values of the washout ratio and M were then determined by calibration of the DOC concentration in runoff and in very low baseflow to values derived from the observed data. The procedure was repeated until the parameter values stabilised (usually after two iterations).

The calibrated values and goodness-of-fit statistics for DOC are given in Table 13.3, along with the average annual flux of DOC produced by the model.

Table 13.3 Calibrated parameters and goodness-of-fit statistics for the DOC model

Catchment	a_{279} gC/m^2/day	b	E_a kJ/gC	k_{slow} fraction per cm	Ratio k_{fast} to k_{slow}	M	E_{NS}	R^2	Modelled DOC flux tC/km^2/year
Glenamong, Ireland	0.039	0.0	4.9	0.019	1.3	n/a	0.49	0.49	17.6
Upper Catchment Lough Leane, Ireland	0.013	0.0	9.2	0.003	7.1*	n/a	0.39	0.39	10.3
Trout Beck, UK	0.041	0.0	0.0	0.019	1.7	n/a	0.34	0.35	14.6
Mustajoki, Finland	0.009	11.0	1.0	0.025	2.0	0.35	−0.84	0.14	4.9
Hedströmmen, Sweden	0.008	0.0	9.5	0.005	2.4	0.44	0.31	0.38	3.4

* inclusion of runoff threshold R_0 of 0.10 cm; in other cases, no threshold applied $R_0 = R_{luDOC}$

In the case of the Upper Catchment of Lough Leane, measurements of DOC are taken at the end of a narrow lake, 170 ha in area and 3 km in length. Retention times are estimated to be about 30 days. This has the effect of smoothing the variation in DOC. The data also show that peak DOC tends to be associated with relatively low values of runoff while high values of runoff are associated with low DOC. This implies substantial dilution at high flows. The effect of the lake was represented in the model by a 30-day moving average and the model parameters, including a runoff threshold R_0, were calibrated simultaneously against the measured data.

Two measures of the goodness-of-fit of the model are given in Table 13.3 – the Nash-Sutcliffe efficiency (E_{NS}) and the coefficient of determination (R^2). The Nash-Sutcliffe efficiency for the DOC model is much less than for the streamflow. While this may indicate limitations in the model representation, there are two other reasons for this. First, is the frequency of the data. Although some shorter records of daily data were available, these often had gaps in the record and the longer-term data were weekly. The second reason is that error, in both the timing and magnitude of flows, is already embedded in the model through the hydrological simulation which means that the quality of the DOC simulation may be compromised. The R^2 values show that between 14 and 49% of the variation in DOC is accounted for by the model. For a water quality model, most of the Nash-Sutcliffe efficiencies are very respectable and the R^2 values are similar to that for the BIM model in the Svartberget catchment in northern Sweden (Taugbøl et al., 1994).

Figure 13.2 shows an example of the model fit for the Glenamong catchment in western Ireland. While producing a good representation of the seasonal variation, the model tends to underestimate the amplitude of the short-term dynamics. This may be due to the use of a daily time step but it also illustrates a general point that, when calibrating on time series, there is a tendency in automatic optimisation to produce rather smoothed simulated series if there are timing errors either in storm events or on a seasonal basis.

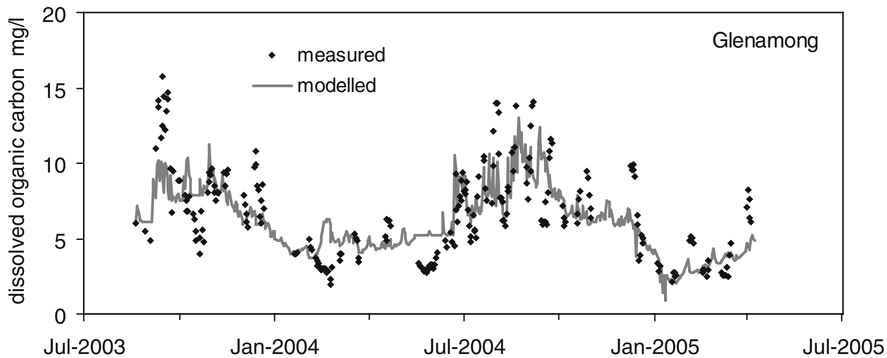

Fig. 13.2 Model simulation and observed values for the Glenamong catchment, Ireland

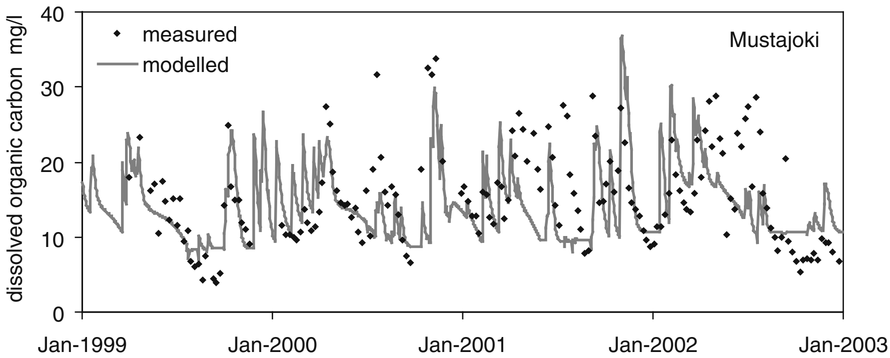

Fig. 13.3 Model simulation and observed values for the Mustajoki catchment, Finland

Figure 13.3 shows an example of the model fit for the Mustajoki catchment in Finland. The overall fit to the full data series is relatively poor (Table 13.3). However, when shorter periods are considered (without re-calibration), some years are seen to perform fairly well. For example, the Nash-Sutcliffe efficiency for 1999–2000 is 0.22 and Fig. 13.3 shows that all the main features (snowmelt peaks and autumn flushes) in the observed DOC concentration for these two years are reasonably well captured in the simulation. However, the model fit in 2001–2002 is relatively poor. One of the reasons for this is that the hydrology model does not simulate the snowmelt period in these years very well – either in terms of volume or timing of flows. A more general point which applies to many of the time series is that some individual years may perform quite poorly while the rest of the time series is quite reasonable. This implies that some catchment dynamics or DOC-related processes are not adequately represented in the model, suggesting that other formulations or alternative model structures might effectively be explored.

In the case of Hedströmmen, DOC data were available from hourly CDOM data for the period September 2003 to December 2005. These were quite anomalous years with 2003 being very dry and 2004 being very wet. This gave some problems in the calibration with unrealistic values of internal variables and uncertain model parameter values. Consequently, long-term monthly TOC data from the neighbouring catchment of Kåfalla were used to estimate the variation in DOC at Hedströmmen for the period 1987–2003. This constrained the calibration to more sensible results and, as shown in Fig. 13.4, gave a good fit to both the short-term frequent measurements (E_{NS} 0.31) and, after an initial spin-up period, to the long-term monthly data (E_{NS} 0.48 for the period 1992–2005). While only 219 estimated values were used alongside the 785 measured values, this dual fit to both long-term data and short-term dynamics, including both very wet and very dry years, gives greater confidence in the use of the model for long-term simulation and shows the value of additional data in constraining model parameters.

When considering the application of models to the assessment of the impacts of climate change, it is also important to assess the calibrated parameter values in

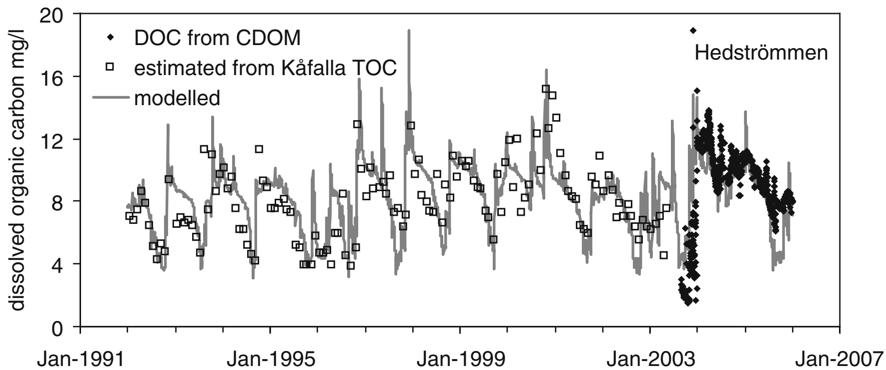

Fig. 13.4 Model simulation and observed values for the Hedströmmen catchment, Sweden

terms of their validity and their likely applicability both historically and into the future. Parameters were calibrated on the observed time series of DOC concentration at the catchment outlet. They show a wide range of values across the catchments considered (Table 13.3). The parameters describing decomposition (a_{279}, b, E_a) may be compared with values from experimental data quoted in the literature. Decomposition of organic matter is more often quoted in terms of Q_{10} values which, for a known temperature, may be converted to activation energy. For peat soils, the range of activation energy values, based on CO_2 measurements, is variously given as 3.6–10.6 kJ/gC for Scottish peats (Chapman and Thurlow, 1996; Chapman and Thurlow, 1998), 5–6 kJ/gC for some Irish peats (Byrne et al., 2001) and 0.5–9.1 kJ/gC for Finnish peats (Silvola et al., 1996). Most of these values are somewhat greater than the available values based on DOC measurements which range from 2.3–4.8 kJ/gC (Moore and Dalva, 2001; Clark, 2005) and are perhaps more likely to represent the net rates of DOC production. With the exception of Trout Beck, the calibrated values of the activation energy fall within the wider range of the quoted variation, with the values for the Upper Catchment and for Hedströmmen being near the upper end of this range.

In the case of dependence on soil moisture, as seen through the b parameter, the evidence from the literature suggests that aerobic decomposition rates are between 3 and 5.5 times anaerobic decomposition rates (Silvola et al., 1996; Evans et al., 2006). When applied at a given temperature, the calibrated parameter value for Mustajoki gives a ratio of aerobic to anaerobic decomposition of about 3.8 i.e. in the middle of the quoted range. In all the other catchments, the calibrated parameters show no dependence of decomposition on soil moisture. The fact that temperature and soil moisture are highly correlated has a confounding effect in the calibration such that separate values for temperature and soil moisture dependence may not be clearly identified (Davidson and Janssens, 2006). To illustrate the two effects described in the model, Fig. 13.5 shows the simulated DOC production over two annual cycles for the two northern catchments. Both have a similar calibrated anaerobic decomposition rate at the reference temperature of 6°C. Hedströmmen, shown

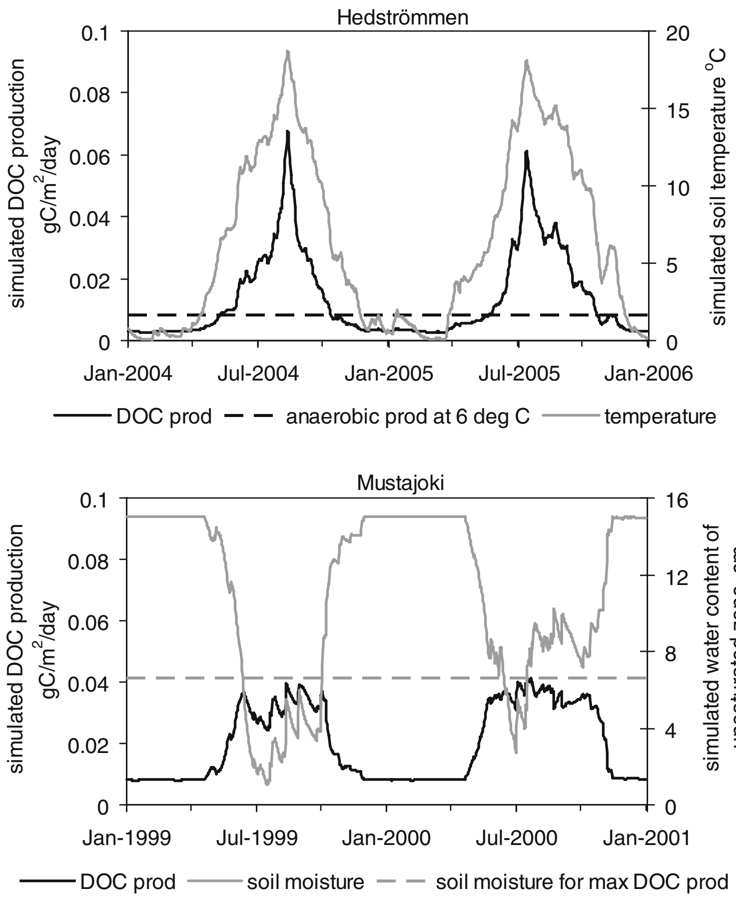

Fig. 13.5 Modelled DOC production under temperature and soil moisture dependence for the two northern catchments of Hedströmmen (*top*) and Mustajoki (*bottom*)

at the top, has a calibrated high dependence on soil temperature which can be clearly seen in the graph with lower simulated DOC production during the winter and over a tenfold increase in the summer. The calibrated parameters for Mustajoki indicate minimal dependence on temperature but a strong dependence on soil moisture. This is shown at the bottom of Fig. 13.5. The model assumes that the decomposition rate peaks at a soil moisture about 70% of saturation (Mitchell and McDonald, 1992) and, for Mustajoki, this corresponds to a soil moisture content of the unsaturated zone, as defined in the GWLF model, of 6.6 cm. Thus, while there is a steep increase in the simulated DOC production as soil moisture drops below saturation, it flattens off and even declines for higher soil moisture deficits, producing the flattened upper parts of the graph and similar patterns in the two years despite the much more prolonged dry summer in 1999. Overall, Fig. 13.5 shows that, while the general seasonal variation in the simulated DOC production is similar, it differs in shape and detail depending on whether it is driven by temperature or soil moisture. It is also worth noting that, in the wetter western catchments, soil moisture deficits are generally smaller so, under current climate, the simulated DOC production would only show an increase with decreasing soil moisture, and the seasonal variation would, therefore, be more similar in shape to the temperature-driven curve. This implies that it is only possible to apportion the effect of these two influences through experimental evidence. It also implies that examining the impacts of climate change through a single set of parameter values for a single catchment may give misleading results dependent on the relative changes in temperature and soil moisture under future climate scenarios.

Taking this discussion further, the case of Trout Beck is instructive insofar as there are experimental data on decomposition rates available for the site (Clark, 2005). These measurements come from highly controlled experiments on the peats in the Trout Beck catchment and give average values of E_a, b and a_{279} of 4.0 kJ/gC, 9 and 0.019 gC/m^2/day respectively. These values indicate that decomposition is dependent on both temperature and soil moisture and do not agree with the calibrated values given in Table 13.3. When the measured values are applied in the model and the other model parameters recalibrated, the seasonality of DOC concentrations is not correctly simulated – high DOC concentrations continue much further into the winter and early spring and the amplitude of the seasonal variation is reduced as the remaining parameters try to minimise the overall sum of squared errors. This mismatch in timing and the associated smoothing of the model response is reflected in a much reduced Nash-Sutcliffe efficiency of 0.18. The issue of parameter values can also be approached through an uncertainty analysis (e.g. Beven and Freer, 2001; Pohlert et al., 2007). Initial results for Trout Beck show that there are many parameters sets which do almost as well as the calibrated values in terms of fit (i.e. $E_{NS} > 0.3$) and these have ranges of E_a of 0.06–4.7 kJ/gC and b of 0.03–2.7. While this goes some way towards reconciling model parameter values with the experimental data, the value of the b parameter remains relatively small and there are no parameter sets which combine high dependence on temperature with high dependence on soil moisture. Another avenue to explore, then, is the uncertainty in model structure. The seasonality mismatch noted above may well be improved by using two DOC stores – one associated with washout in runoff which becomes exhausted during autumn storms (cf. Worrall et al., 2002) and one associated with subsurface washout. However, any improvement in model performance would need to be balanced against the increased number of parameters and assumptions required by the model.

Looking at the other parameter values (Table 13.3), anaerobic production rates are calibrated to give total decomposition roughly in balance with the DOC flux from the catchment. They, therefore, vary across the sites as a function of the other decomposition parameters e.g. the value for Trout Beck is comparatively high as decomposition is constant throughout the year; the value for Hedströmmen is comparatively low as there is a large seasonal variation on top of the background

Table 13.4 Simulated partitioning of DOC fluxes and concentrations between hydrological pathways

	Percentage of water flux			Percentage of DOC flux			average DOC concentration mg/l	
Catchment	Runoff	Fast sub-surface	Slow sub-surface	Runoff	Fast sub-surface	Slow sub-surface	Runoff	Sub-surface
Glenamong	55	33	12	62	27	11	6.2	6.0
Ireland	*51*			*59*			*9.4*	*6.8*
Upper								
Catchment	39	49	13	46	42	12	5.0	4.7
Lough Leane	*38*			*40*			*4.8*	*4.8*
Ireland								
Trout Beck	65	30	5	75	20	4	12.2	7.5
UK	*65*			*73*			*12.8*	*8.7*
Mustajoki	29	47	23	48	41	11	28.6	11.0
Finland	*31*			*44*			*25.6*	*14.6*
Hedströmmen	18	69	13	24	70	6	8.4	8.5
Sweden	*16*			*19*			*9.8*	*9.7*

Estimated values from data shown in italics

decomposition rate caused by the calibrated high dependence on temperature. The parameters controlling washout of DOC are a function of the catchment hydrological response and partitioning between pathways should reflect the observed concentrations in streamflow and baseflow.

Table 13.4 shows the water and DOC fluxes as well as the DOC concentrations in the different hydrological pathways for each of the catchments. For comparison, estimates based on the available data are shown in italics. The percentage of the water flux is based on daily data and the baseflow separation method of Arnold and Allen (1999). For the DOC, estimates are much less certain due to the limited number of samples and the additional assumptions made. The DOC concentration in baseflow is assumed to be best described by the average DOC in those samples when the flow is only made up of baseflow. The concentration of DOC in runoff is derived from a mixing model using the baseflow-separated flows and a linear interpolation of the baseflow-only concentrations. Fluxes of DOC are calculated as a flow-weighted flux given by the ratio estimator (Cooper and Watts, 2002). In general, the simulation model is seen to perform well against these approximate values. In the case of Trout Beck, the partitioning of DOC concentrations and fluxes is also consistent with the independent analysis of Worrall et al. (2006) which shows that the majority of DOC is sourced from shallow soil water.

The 'available' DOC store generated within the model (see Fig. 13.1) can be compared with measured pore water concentrations by making assumptions about the total soil moisture content. Model values are generally in line with published data (e.g. Aitkenhead-Peterson et al., 2003; Starr and Ukonmaanaho, 2004; Clark et al., 2005). The variation in this store may also be compared with published values of total soil carbon. Taken across all the sites, the 'available' DOC store has an annual variation of about 5 gC/m^2. This can rise to about 10 gC/m^2 if the inter-annual variation is taken into account. By comparison, published values of total soil carbon are 3,060 gC/m^2 for the top 40 cm of upland forest soils in Finland (Ilvesniemi et al., 2002) and over 20,000 gC/m^2 for the top 1 m of upland blanket peats in the UK (Bradley et al., 2005). Even if most of the carbon leached from blanket peats is from the top 10 cm, these figures lend support to the assumption that net DOC production is not limited by the total carbon store. Indeed, taken as a whole, annual leaching of DOC across Finland is estimated to be <0.1% of the total carbon store (Arvola et al., 2004).

Along with the goodness-of-fit to observed DOC concentrations, this comparison of calibrated values and internal model variables with published figures provides additional confidence in the use of the model. It is also clear from the above discussion that the values taken by the model parameters will have a profound effect on the projected impacts of climate change. Here, prior to a full exploration of parameter uncertainty, a single set of optimised parameter values has been assumed and the variation in the projected response to climate change is illustrated by the different catchments in each of the two regions – almost as a sensitivity analysis rather than a prescribed projection for the particular catchments simulated. For example, the results from the calibrated parameters for Trout Beck are included to illustrate the impact under climate change of changes in hydrology on DOC, in the absence of changes in decomposition, as a limiting case for western catchments.

13.4 Modelling the Impacts of Climate Change on Dissolved Organic Carbon

In order to provide some indication of the potential impact of climate change on dissolved organic carbon, we have taken the pragmatic step of assuming that the calibrated parameters and the model structure are appropriate for simulating dissolved organic carbon in the future. Just as we have rightly flagged uncertainty in model parameters, uncertainty in future climate projections also needs to be considered. Here, three main components of uncertainty may be identified: the chosen greenhouse gas scenario; the climate model or combination of Global (GCM) and Regional Climate Model (RCM) used; and the individual realisation of daily precipitation and temperature downscaled to each catchment. The climate scenarios applied were described in Chapter 2 and are summarised in Table 13.5 .

For each climate scenario, downscaling to the individual catchment was provided either by a weather generator (Jones and Salmon, 1995; Watts et al., 2004) or by the delta change, sometimes referred to as the incremental scenario, approach (Hay et al., 2000; Andreasson et al., 2004). The advantage of the weather generator was that it not only enabled downscaling to the catchment but could also be

Table 13.5 The future climate scenarios applied

Abbreviation	GCM	RCM	Greenhouse gas scenario
H A2	HadAM3h	RCAO	A2
H B2	HadAM3h	RCAO	B2
E A2	ECHAM4/OPYC3	RCAO	A2
E B2	ECHAM4/OPYC3	RCAO	B2
Had A2	HadAM3p	HadRM3p	A2
Had B2	HadAM3p	HadRM3p	B2

used to produce multiple realisations of weather so that some measure of variability could be derived. This is particularly important for DOC as temporal dependence on the *sequence* of wet and dry years seems to be important (Naden and McDonald, 1989; Watts et al., 2001). Accordingly, a resampling method was devised to generate alternative weather sequences when using the delta change approach. The resampling works on a monthly basis and simply chooses a year at random to provide that month's daily precipitation and temperature data. Thus, the method preserves the integrity of the weather within a month but may give some discontinuity across the month ends. A comparison of monthly temperature and precipitation shows that, for the Upper Catchment of Lough Leane, the two methods produce a similar degree of variability. The delta change method was employed in the two northern catchments where weather generator data were not available. One hundred realisations of 30 years of daily precipitation and temperature values were generated as input to the hydrology and DOC model. The control run was based on long-term observed meteorology, while the scenarios were derived by perturbing the control by the differences in the RCM output for each control-scenario pair (Chapter 2).

When applying the climate change scenarios, the growing season was also adjusted using the method based on daily mean temperature developed by Mitchell and Hulme (2002). This method has been shown to correspond well with data from Europe and northern latitudes and gave increases of between 25 and 97 days in the growing season dependent on the catchment and the climate scenario. The results of the climate change simulations have been expressed in terms of the variability in the annual and monthly mean DOC concentrations and fluxes. Details of hydrological changes were described in Chapter 3.

13.4.1 Changes in Annual Mean DOC

Figure 13.6 shows the projected annual mean DOC concentration for 2071–2100 in the western catchments. In each case, the median value of the simulated annual mean over the 100 realisations of 30 years is shown by the dash; the interquartile range is shown by the box and the whiskers give the 5 and 95 percentiles. Present-day

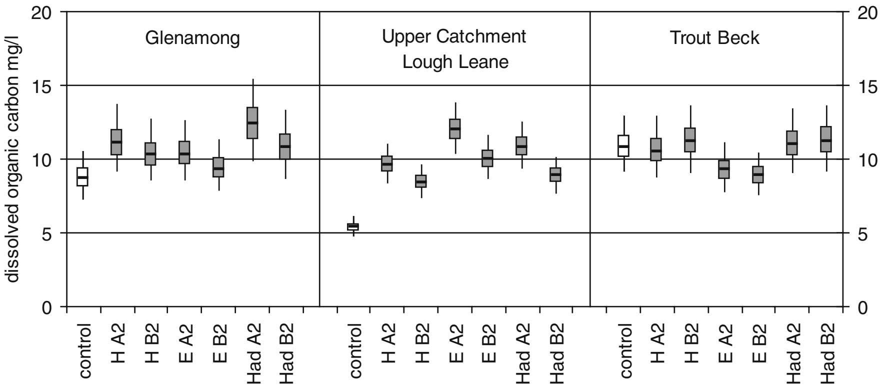

Fig. 13.6 Projected annual mean DOC concentration (2071–2100) in western catchments

conditions are indicated by the control runs, shown in white, to the left of the climate change scenarios. In all cases, measured values of the annual mean DOC concentration fall within the variability of the control runs.

The climate change projected for the western catchments, although varying in detail, shows a similar pattern with an increase in temperature of between 1.5°C and 4°C and a shift towards wetter winters (Oct–Mar) and drier summers (Jun–Sep). This leads to large increases in summer soil moisture deficits which extend well into the autumn. Although there are quantitative differences between projected changes in temperature and precipitation for the individual catchments, the different DOC projections given in Fig. 13.6 largely relate to the different calibrated DOC model parameters. The calibrated parameters for both the Glenamong and the Upper Catchment of Lough Leane indicate that decomposition is highly temperature dependent and both catchments show a significant increase in DOC concentration (Kolmogorov-Smirnov test on 1 in 10 randomly sampled values; $n = 300$, $p < 0.001$) under all the climate scenarios. However, the calibrated activation energy for the Upper Catchment of Lough Leane is approximately twice that for the Glenamong and this accounts for the large difference between the responses of the two Irish catchments. Annual mean concentrations in the Glenamong are projected to increase by 20% from a median value of 8.7 (5 and 95 percentiles: 7.3–10.5) mg/l to 10.5 (5 and 95 percentiles: 8.5–13.7) mg/l when considered across all climate scenarios. In the Upper Catchment of Lough Leane, the projected increases are from a median of 5.4 (5 and 95 percentiles: 4.8–6.0) mg/l to 8.9 (5 and 95 percentiles: 7.9 and 12.6) mg/l, an increase of 65%. Although this increase is substantial, the lower projected values for the Upper Catchment are within the present-day range of values for the Glenamong. The high projected DOC concentrations would have ecological implications for Lough Leane and other lakes in the Leane catchment (see Chapter 12). While increases in DOC availability can lead to higher rates of primary production, particularly in small sheltered lakes (Jones, 1992), high water colour has been associated with light limitation of phytoplankton photosynthesis in some larger, more exposed Irish lakes such as Lough Leane (Jewson and Taylor, 1978; Foy et al., 1993; Girvan and Foy, 2006). In addition, Lough Guitane, a small lake in the peat area of the Leane catchment, is the water source for about 40,000 people. The projected increases would require additional investment in water treatment. Difficulties in treating water could also be exacerbated due to the increased variability in concentrations, particularly in the autumn (see below).

In the case of Trout Beck, the calibrated parameters showed no dependence on either temperature or soil moisture and the projected change in DOC concentration is purely due to hydrological change rather than to any change in peat decomposition. Most of the climate scenarios give similar values to the present-day. The exceptions to this are the climate scenarios produced by the E A2 and E B2 scenarios which are generally warmer and wetter and, thus, project a significant (Kolmogorov-Smirnov test on 1 in 10 randomly sampled values; $n = 300$, $p < 0.001$) decrease in DOC concentration from 10.8 mg/l to around 9 mg/l for this site under these parameters. It should be stressed that the results for Trout Beck are simply included to illustrate the lower limit of expected changes in western catchments – experiments on peat soils from the Trout Beck catchment show that their decomposition is dependent on both temperature and soil moisture.

Figure 13.7 shows a similar plot of the projected (2071–2100) annual mean DOC fluxes. Again, there are significant increases projected for both the Irish catchments from a median of 18 (5 and 95 percentiles: 15–21) tC/km^2/year to 23 (5 and 95 percentiles: 23–28) tC/km^2/year across all scenarios for the Glenamong and from a median of 9 (5 and 95 percentiles: 7–11) tC/km^2/year to 17 (5 and 95 percentiles: 12–23) tC/km^2/year across all scenarios for the Upper Catchment of Lough Leane. There is no significant change in flux projected for Trout Beck. The large differences between catchments arise from differences in the calibrated DOC model parameters rather than differences in the climate change scenarios for the different sites. However, within each catchment, or set of model parameters, the variation between climate scenarios can be substantial. Based on the median values, the annual mean DOC flux is projected to increase by between 14 and 40% in the Glenamong and between 54 and 127% in the Upper Catchment of Lough Leane. Figure 13.7 shows that the pattern is similar across the six scenarios for these two sites with the smallest increase projected with the H B2 scenario and the largest increase by the E A2 scenario. The greater range of variation in the projected DOC flux for the Upper Catchment of Lough Leane arises from the higher calibrated dependence of decomposition on temperature. Within each pair of climate scenarios, the increase in DOC flux is always higher under the A2 scenario compared with the B2 scenario as expected. However, for any given emission scenario, the difference in projected flux from the different climate models is greater than that from the different emissions scenarios applied to any single climate model.

Turning to the northern catchments, the projected changes in climate are again fairly similar with projected annual mean temperature increases of between 2.9°C and 5°C. Annual precipitation is also projected to increase, with substantial increases in winter precipitation (Dec–May). The main implication of this is much

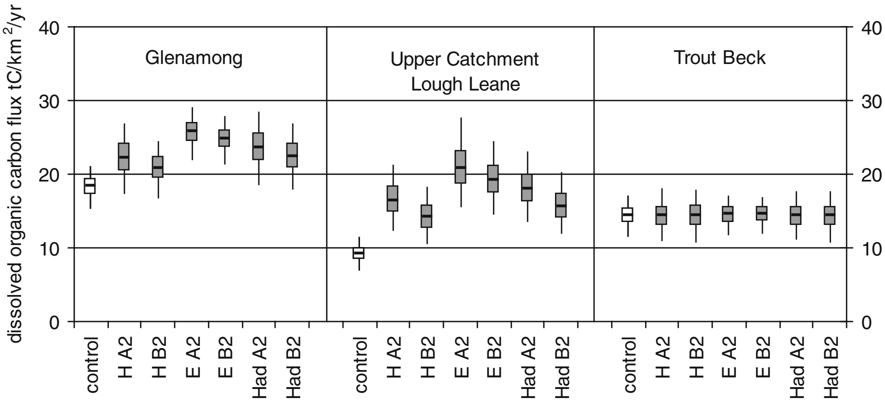

Fig. 13.7 Projected annual mean DOC flux (2071–2100) in western catchments

greater streamflow in the winter months and the disappearance of the snowmelt peak, causing a fundamental change in the seasonality of both the hydrology (see Chapter 3) and the DOC (see below). Figure 13.8 shows the projected values for the annual mean DOC concentrations. Again, measured values of the annual mean fall within the variability of the control runs.

In Mustajoki, concentrations taken across all future scenarios are projected to be in a similar range (5 and 95 percentiles 10.3–21.1 mg/l) to present day (5 and 95 percentiles across all scenarios 9.7–20.2 mg/l). However, there is considerable variation between the results from the different climate models with the E A2 scenario giving a small decrease of 1.2 mg/l and the Had A2 scenario giving a sizable increase of 3.5 mg/l. It is noteworthy that the variability between years is much greater in Mustajoki than in other catchments. This may be due to the higher variability of

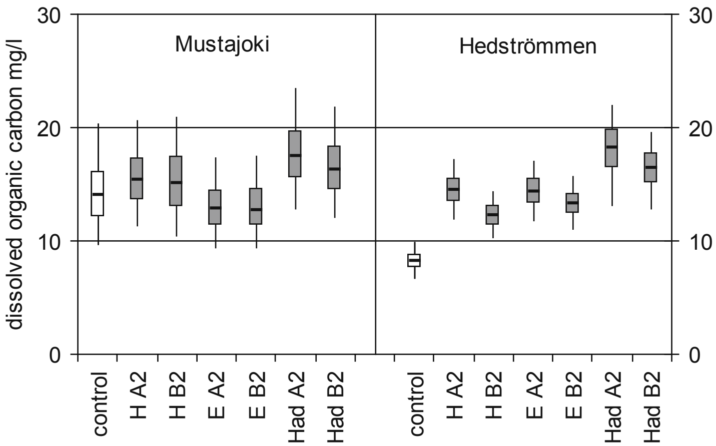

Fig. 13.8 Projected annual mean DOC concentration (2071–2100) in northern catchments

precipitation in more continental areas. It could also arise from the calibrated DOC model parameters which for Mustajoki, distinct from the other catchments, show a strong dependence of decomposition on soil moisture, with only a slight dependence on temperature.

In contrast, Hedströmmen shows a significant increase in DOC concentrations for all scenarios (Kolmogorov-Smirnov test on 1 in 10 randomly sampled values; $n = 300$, $p < 0.001$). This follows from the high dependence of decomposition on temperature given by the calibrated parameter values for this catchment. The projected increase, taken across all scenarios, is from 8.2 (5 and 95 percentiles: 6.8–9.8) mg/l to 14.4 (5 and 95 percentiles across all scenarios: 11.1–19.7) mg/l. Similar results are found under the H A2 and E A2 scenarios (K-S test $p > 0.1$) but the Had A2 scenario gives a significantly (K-S test $p < 0.001$) larger increase in DOC. This is due to higher summer temperatures and lower autumn precipitation compared with the other climate models. Hedströmmen is one of the catchments feeding into Lake Mälaren which provides the water supply for Stockholm. The large projected increases in DOC concentration are likely to have a significant impact on the treatment costs of this potable water supply.

Figure 13.9 shows the projected DOC fluxes for the northern catchments. For Mustajoki, there is a small but significant (Kolmogorov-Smirnov test on 1 in 10 randomly sampled values; $n = 300$, $p < 0.001$) increase in annual flux for all scenarios from 4.8 (5 and 95 percentiles: 2.9–7.4) tC/km^2/year to 5.6 (5 and 95 percentiles: 3.5–8.1) tC/km^2/year. A similar increase in the median value of the annual mean flux is found across all the different climate scenarios, ranging between 13 and 21%.

For Hedströmmen, a substantial increase in flux is projected across all scenarios from 3.2 (5 and 95 percentiles: 1.9–4.9) tC/km^2/year to 5.8 (5 and 95 percentiles: 3.2–9.2) tC/km^2/year. In this case, due to the high calibrated dependence of decomposition on temperature, the increase is both more substantial and more variable

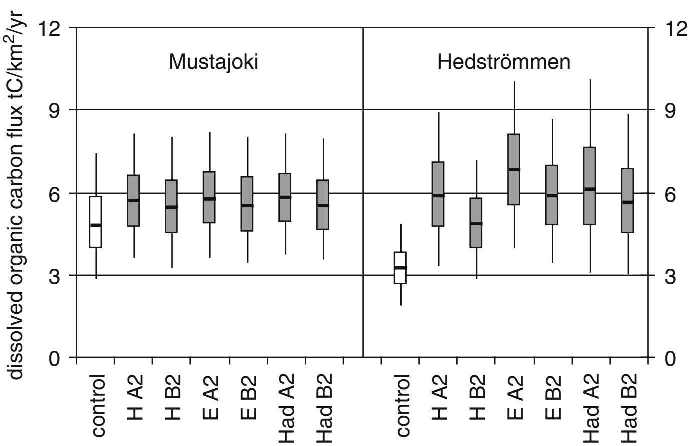

Fig. 13.9 Projected annual mean DOC flux (2071–2100) in northern catchments

across the different scenarios. The increase ranges from 35% in the H B2 scenario to 96% in the E A2 scenario. While the highest increases in DOC concentration were projected under the Had A2 scenario, the largest DOC fluxes are projected by the E A2 scenario. Under this scenario, the near doubling of DOC flux into Lake Mälaren would have a large impact on the attenuation of photosynthetically-active radiation and lead to changes in the depth of the euphotic zone. Short-term variations in the import of coloured organic matter (gelbstoff) to Lake Mälaren during periods of high river discharge have already been shown to be an important factor influencing the depth of the euphotic zone (Pierson et al., 2003). Hedströmmen flows into Galten, the westernmost basin of Lake Mälaren. Galten is a eutrophic shallow basin where phytoplankton species richness is high due to infrequent, short periods of thermal stratification (Willén, 2001) and the migration of species from a high number of contributing lakes in the catchment. Abundant nutrients, in combination with unstable stratification, may lead to a shift in phytoplankton species composition favouring colonial species of buoyant cyanobacteria (Reynolds and Walsby, 1975; Pierson et al., 1994). While changes in the light attenuation are most likely to influence primary production, catchment sources of DOC can also shift lake metabolism and the structure of biotic communities (Tranvik, 1992; Jansson et al., 2000). Additionally, an increased subsidy of carbon from the catchment has important implications for release of CO_2 from the lake, as it becomes an increasing source for atmospheric CO_2 with the increasing concentration of coloured organic matter in the lake water (Hope et al., 1996; Sobek et al., 2003).

13.4.2 Changes in Seasonal Patterns

The seasonal pattern of concentrations and fluxes is perhaps of even more significance than the annual mean for both the ecological consequences and for the treatment costs of potable water. To illustrate changes in seasonal pattern, just one of the climate change scenarios has been selected – the H A2 scenario which has a central position in the overall range of results – and projections are shown for two example catchments representing western and northern Europe respectively.

Figure 13.10 shows monthly averages of both concentration and flux for the Glenamong catchment in Ireland under both the control and the H A2 climate change scenario for the model parameter values given in Table 13.3. A pair of results is given for each month with the control in white on the left and the climate change projection in grey on the right. While annual mean concentrations show an increase of around 2.4 mg/l for the H A2 scenario, Fig. 13.10 shows that large increases in concentration are projected in the late autumn and early winter. This results from increased decomposition during the summer, due to higher temperatures, coupled with enhanced washout by higher precipitation in autumn and winter. Concentrations in October are projected to rise from 12 (5 and 95 percentiles: 8.4–16.3) mg/l to 20 (13.5–24.3) mg/l; in November from 10 (6.6–14.1) mg/l to 18 (11.3–24.3) mg/l and in December from 8 (5.3–11.7) mg/l to 13 (7.8–20.2) mg/l. This large increase

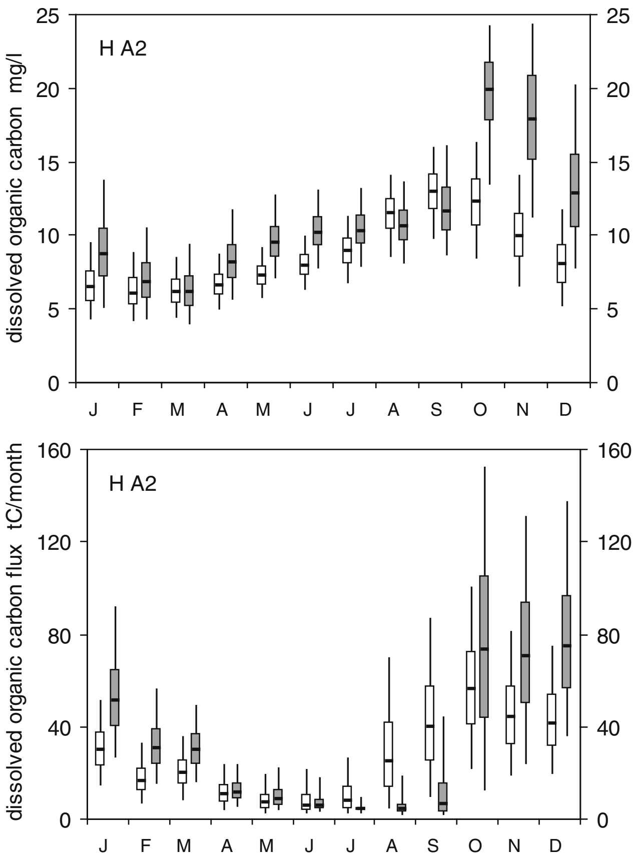

Fig. 13.10 Seasonal change in both DOC concentration and flux for the Glenamong shown as monthly mean values. Median shown by *dash*; interquartile range by *box*; 5 and 95 percentiles by whiskers; control in *white*; future projection (2071–2100) in *grey*

of between 60 and 80% in the monthly mean DOC concentrations, and hence in water colour, will have significant implications for the treatment costs of water for public consumption.

The main features of the seasonal pattern in DOC fluxes projected by the model for the H A2 scenario (Fig. 13.10) are a significant decrease in flux in August and September as well as a significant increase in flux between October and March. The reduced flux at the end of the summer is due to the large soil moisture deficit built up during the summer. This causes delayed re-wetting of the catchment and substantially reduced streamflow in the late summer and early autumn. The winter increase in DOC flux is partly due to the higher streamflow caused by higher rainfall and partly due to the higher DOC concentrations. The Glenamong catchment drains into Lough Feeagh which is an important fish habitat, especially for the freshwater stages of salmon and sea trout. In addition to impacts on phytoplankton and zooplankton productivity, reduced light intensity also lowers the reactive distance of fish and their ability for size-selective predation, while decreased oxygen levels in the hypolimnia of humic lakes can provide a refuge for prey species which are less sensitive to oxygen availability than fish (Wissel et al., 2003). The effects of higher DOC concentrations in Lough Feeagh are, therefore, expected to be a reduction in primary productivity and in food availability and habitat for fish species.

Figure 13.11 shows the monthly variation in DOC concentrations and fluxes for Mustajoki under both the control and H A2 climate change scenario for the calibrated model parameters given in Table 13.3. In the case of DOC concentrations, the increase in the annual mean of 1.4 mg/l under the H A2 scenario is mainly seen

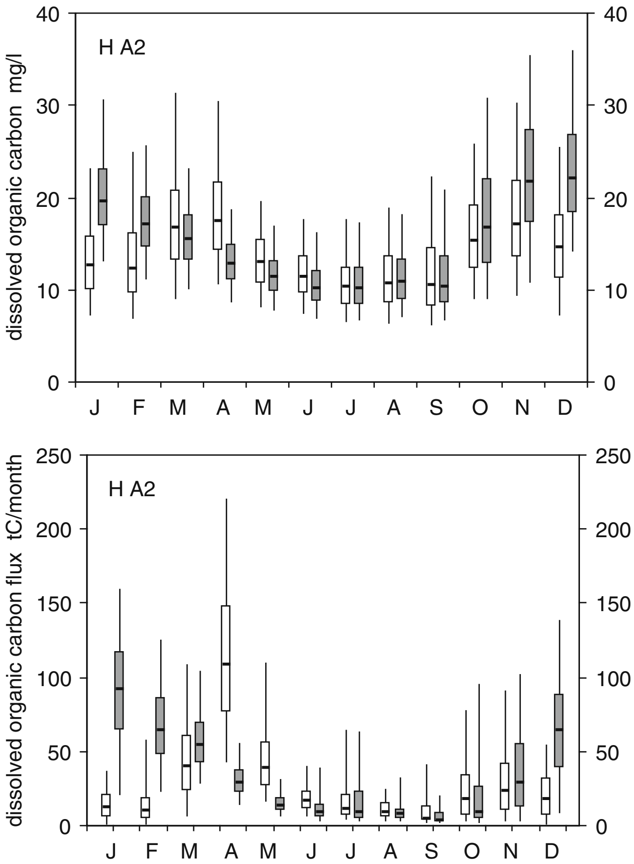

Fig. 13.11 Seasonal change in both DOC concentration and flux for Mustajoki shown as monthly mean values. Median shown by *dash*; interquartile range by *box*; 5 and 95 percentiles by whiskers; control in *white*; future projection (2071–2100) in *grey*

to occur in the winter months – particularly November to February with significant increases in concentration from 17 to 22 mg/l in November, 15 to 22 mg/l in December, 13 to 20 mg/l in January and 12 to 17 mg/l in February. This increase is partly due to increased decomposition but is also the result of higher winter temperatures, the loss of snow cover, higher winter rains and hence higher streamflows. There is a significant decrease in concentration in April from 17 to 13 mg/l. This is associated with the loss of the snowmelt peak and the continued flushing of the peat soils through the winter period.

These hydrological effects are even more striking in terms of the DOC flux. Whereas the seasonal pattern is currently dominated by the snowmelt peak in April/May with a secondary peak in the autumn (October–December), the future projection is for the seasonal flux to be dominated by a winter peak extending from December to February. Fluxes in January are projected to increase from 11.7 (5 and 95 percentiles: 1.4–36.2) tC/month to 92.1 (21.5–158.9) tC/month while the flux in April is projected to decrease from 108.2 (42.8–219.8) tC/month to 28.9 (14.5–55.6) tC/month. In terms of the ecological implications for Lake Pääjärvi, it is expected that these large seasonal changes will help to mitigate any impact of the increase in the annual flux. For example, the largest increases in flux are projected to occur in winter when the temperature of the inflowing water is very low. Given the inverse thermal stratification in the lake, at least part of the inflowing water is expected to flow across the surface direct to the outflow without complete mixing with the main body of lake water.

While the seasonal changes in flux projected for Mustajoki appear dramatic, it should also be remembered that the calibrated parameter values give only a very low dependence of decomposition on temperature and that the projected changes, therefore, largely reflect the fundamental changes projected in the hydrology of northern catchments. With a high dependence of decomposition on temperature and climate change scenarios suggesting temperature increases of 2.9–5°C, the projected seasonal variation would be considerably enhanced. The example of Hedströmmen, with its high calibrated dependence on temperature, is perhaps illustrative of an upper limit to the expected change. Here, projections are for enhanced DOC concentrations throughout the year but particularly in the winter months. As a result of this and the seasonal hydrological changes, DOC fluxes in January to March are projected to increase by 160–290%. Smaller changes are projected for the rest of the year with an 18% increase in April/May, a 35% decrease in summer (July–September) and a 58% decrease in October due to lower summer streamflow and delayed re-wetting of the catchment following large summer soil moisture deficits.

13.5 Discussion

This chapter has shown the range of changes in DOC concentrations and fluxes projected under the six climate change scenarios described in Chapter 2 using the combined GWLF hydrological model, described in Chapter 3, and the DOC model described here. Based on the example catchments from both Western Europe (Ireland and the UK) and Northern Europe (Finland and Sweden), the general picture is for an increase in both DOC concentrations and fluxes. For some catchments, the projected increases are substantial with implications for both lake ecology and for the costs of treating potable water.

In summary, taken across all the climate scenarios, the projected changes for the Irish catchments are a 20% increase in annual mean concentration and 18% increase in annual mean flux for the Glenamong and a 65% increase in annual mean concentration and 89% increase in annual mean flux for the Upper Catchment of Lough Leane. For the northern catchments, the projected changes are a 7% increase in annual mean concentration and 16% increase in annual mean flux for Mustajoki and a 74% increase in annual mean concentration and 79% increase in annual mean flux for Hedströmmen. Projected seasonal changes are even more profound. Concentrations in the Glenamong are projected to increase by as much as 80% in the autumn. In northern catchments, fundamental changes to the hydrology and the loss of the snowmelt peak, lead to large increases in DOC flux in the winter which may be associated with large reductions in DOC flux in the spring. However, as with all projections into the future, these results are entirely dependent on the climate change scenarios used, the validity of the calibrated model parameter values, and the assumptions of the model and its ability to represent relevant processes. Each of these elements now needs to be scrutinised in terms of its relative importance and bearing on the results.

Within CLIME, by using results from different climate models, we have been able to show something of the impact on DOC of the uncertainty in climate change scenarios – not only in terms of the emissions scenario but also the climate forcing or GCM/RCM combination. Investigation into the climate forcing has also been extended through the use of a weather generator or resampling technique to look at the impact of multiple realisations of daily precipitation and temperature series in order to quantify what might be termed natural variability within any one scenario. Although a fairly limited range of climate models were used, the results indicate that, in most cases, the differences between the climate models is greater than the differences between the two emissions scenarios run through the same climate model. Furthermore, with the exception of those catchments with calibrated DOC parameters which show a very high dependence of decomposition on temperature, the extent of the differences in the projected median values from all the future climate scenarios tends to be less than the range of the inter-annual variability (as expressed by the 5 and 95 percentile values generated from the 100 realisations of the 30-year simulations). This has also been found for other variables in other studies (e.g. Hulme et al., 1999; Arnell, 2003). It highlights the importance of reporting projected change in the context of both climate uncertainty and natural variability.

Another point to come out of this study is the large differences in the calibrated DOC model parameters across the catchments. This is particularly important with regard to the dependence of organic matter decomposition on temperature and soil moisture and its propagation through to climate change impacts on DOC levels. Although the majority of calibrated values has been shown to fall within the wider range of values quoted in the literature, it has also become clear that it is impossible to separate out temperature and soil moisture effects through calibration. We also know that there are many different sets of parameter values which could give almost as good a fit to the observed data and, in future, a formal approach to the issue of model parameter uncertainty needs to be taken and followed through to the quantification of uncertainty in future projections (cf. Cameron et al., 2000). Only a single set of optimised parameter values has been used here but, if we consider the catchments in a region as representing an ensemble of possibilities, then it is clear that the variation in the projected DOC concentrations and fluxes arising from the different calibrated DOC (and hydrology) model parameters is far greater than the variation that arises from the different climate change scenarios. The implication of this is that, if we wish to reduce uncertainty in future projections of DOC, effort needs to be placed on the identification of model parameter values through field and experimental measurements.

The issue of uncertainty in the model structure has already been highlighted by the example of Trout Beck, with the failure to reconcile calibrated parameters with those determined experimentally. This discussion can be broadened to consider the model's simplifying assumptions more widely. Fundamental among these are the use of a single carbon pool and the assumption that the production of DOC is not limited by the availability of organic matter. The size of the source pool is important because the model assumes that, under future climate scenarios with higher temperatures and lower soil moisture, DOC production will increase. Given the total carbon store in peat soils and the rates of DOC production, this may not be a problem. However, particularly in forest soils, other workers have proposed that much of the DOC is from fresh plant litter which is a relatively small pool. Support for this idea comes from ^{14}C evidence which suggests that most DOC is relatively young (Schiff et al., 1997; Tipping et al., 2005). There is also a great deal of current work which shows stream DOC to vary in its properties with flow, i.e. at low streamflow material is different from that at high streamflow in terms of its molecular composition, isotopic content and functional properties (e.g. Sharp et al., 2006), suggesting that it is sourced from different pools. It is also clear that the way the hydrology interacts with these different pools varies between catchments and has different spatial expression (Hinton et al., 1998). This implies the need for a spatially-distributed hydrological model and a large number of additional parameters whose values are uncertain. How far it is necessary and effective to describe all this complexity in order to project future levels of DOC at the catchment scale is not really known.

Indeed, it is recognised that DOC is part of a much wider biogeochemical picture which includes net primary productivity (Neff and Hooper, 2002), vegetation response to increasing CO_2 levels (Freeman et al., 2004), recovery from acid deposition (Chapman et al., 2005; Evans et al., 2006), adsorption-desorption reactions (Tipping and Hurley, 1988; Kalbitz et al., 2000), de-gassing of CO_2 from headwater streams (Dawson et al., 2004) as well as land use and land management changes. The relative importance of each of these processes in any one catchment is unknown and the debate about the causes behind the general increase in DOC levels seen in the historical record across both Europe and North America, as discussed in the previous chapter, continues. One thing that is clear is the immense value of long-term datasets. This was illustrated in the example of Hedströmmen where we were able to couple long-term monthly data with more recent high resolution data to constrain model parameter values. In the wider context of future model development and testing, the need for additional data – both good long-term weekly, or preferably daily or sub-daily, time series as well as data describing internal process dynamics – cannot be stressed enough. In the case of historical time series, there is also a need to assemble concurrent meteorological and hydrological data so that model simulations can be tested against long-term datasets in order to establish whether the observed trends or step-changes in DOC can be successfully generated. This hind-casting can then be used as further evidence to support or refute the assumed model structure and associated parameter values used to project future change.

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Chapter 14
The Impact of Variations in the Climate on Seasonal Dynamics of Phytoplankton

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14.1 Introduction

Phytoplankton, an assemblage of suspended, primarily autotrophic single cells and colonies, forms part of the base of the pelagic food chain in lakes. The responses of phytoplankton to anthropogenic pressures frequently provide the most visible indication of a long-term change in water quality. Several attributes related to the growth and composition of phytoplankton, such as their community structure, abundance as well as the frequency and the intensity of blooms, are included as indicators of water quality in the Water Framework Directive. The growth and seasonal succession of phytoplankton is regulated by a variety of external as well as internal factors (Reynolds et al., 1993; Reynolds, 2006). Among the most important external factors are light, temperature, and those associated with the supply of nutrients from point and diffuse sources in the catchment. The internal factors include the residence time of the lakes, the underwater light regime and the mixing characteristics of the water column. The schematic diagram (Fig. 14.1) shows some of the ways in which systematic changes in the climate can modulate these seasonal and inter-annual variations. The effects associated with the projected changes in the rainfall are likely to be most pronounced in small lakes with short residence times (see George et al., 2004 for some examples). In contrast, those connected with the projected changes in irradiance and wind mixing, are likely to be most important in deep, thermally stratified lakes.

In this chapter, we use results acquired from a range of different European lakes to explore the potential effects of climate change on the seasonal development and the composition of phytoplankton. The time-series analysed are amongst the longest available in the region. Some of the lakes studied in CLIME have been sampled at weekly or fortnightly intervals for more than fifty years. These sites also cover a range of lake types from shallow to deep, small to large, and oligotrophic to eutrophic.

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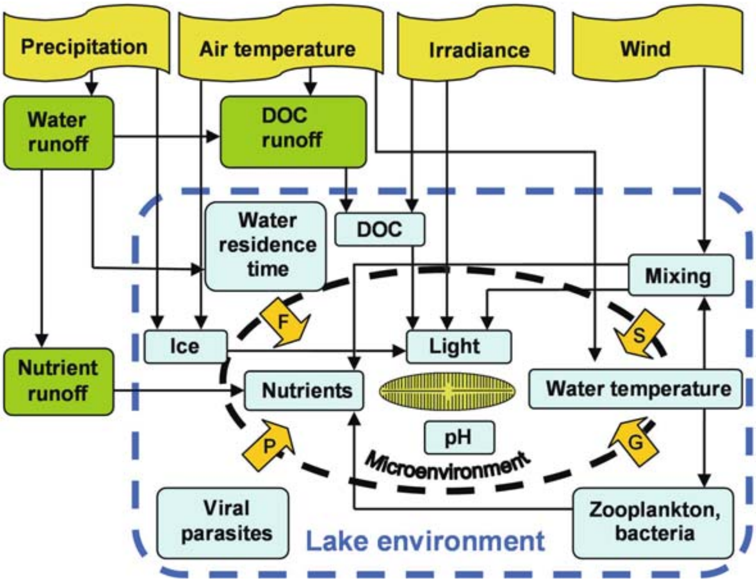

Fig. 14.1 The mechanisms involved in mediating the climatic responses of phytoplankton. The elimination mechanisms are indicated by block arrows: F – flushing by extensive water exchange, S – sedimentation in a stagnant water column, G – grazing by zooplankton, P – parasitism.

Geographically the lakes represent the Nordic, Atlantic, Central European and Alpine climatic and eco-regions. Besides specific CLIME sites, examples from other lakes are included where appropriate. Here, we concentrate our attention on the changes observed over the last 30 years, a period of particularly rapid change in most of the catchments selected for study. Issues that greatly complicate the analysis of long-term climate change impact on lakes, are the complementary trends of eutrophication and re-oligotrophication. These management-related problems can, however, be minimized by using appropriate de-trending techniques and drawing comparisons with model simulations.

14.2 The Impact of Changes in the Weather on the Seasonal Dynamics of Phytoplankton

14.2.1 Winter

In Northern Europe and the Alpine regions of Central Europe, the lakes are usually covered with ice throughout the winter (Fig. 14.2). Cold monomictic lakes in the sub-arctic and high alpine areas experience only a short ice-free period in summer, while lakes in maritime areas in Western Europe seldom freeze during the winter. Most recent climate change scenarios suggest that there will be a marked increase

Fig. 14.2 The distribution of ice-covered lakes in Europe. Areas with mean air temperature in January <0°C (black) are more likely to have lakes with sustained winter ice-cover. Deep lakes that store more heat, may represent exceptions.

in European winter temperatures accompanied by a pronounced extension of the ice-free period (see Chapters 2, 4 and 6, this volume). Large ecological changes have been observed in lakes which have totally lost their winter ice-cover and lakes which were previously covered with ice but have now become temporarily ice-free (Psenner, 2003). Ohlendorf et al. (2000) concluded from their observations on a remote high alpine lake that the mere occurrence of an ice-free period, creating a short productivity pulse, was more important than its duration for preserving a climatic signal in the sedimentary record.

During the ice-cover period, especially when there is thick snow on the ice, photosynthesis becomes severely light-limited and most species of phytoplankton sink in the water column despite some convective mixing. Exceptionally, some motile algae like dinophytes (Arvola and Kankaala, 1989; Weyhenmeyer et al., 1999), cryptophytes (Arvola and Kankaala, 1989; Phillips and Fawley, 2002), chrysophytes (Watson et al., 2001) or flagellated chlorophytes (Arvola and Kankaala, 1989) can concentrate near the surface and give rise to late winter blooms particularly if the ice is clear of snow (Jones, 1991). In Lake Erken, high chlorophyll *a* concentrations were recorded both when the spring peak occurred below the ice and after the ice break-up (Fig. 14.3A). The relative abundance of diatoms was, however, low when the spring peak occurred below the ice (Fig. 14.3B), primarily due to the importance of wind mixing for the large diatoms. Winter diatom blooms, like those dominated by the very small *Stephanocostis chantaicus* in Lake Stechlin, Germany (Scheffler and Padisák, 2000) or by *Aulacoseira baicalensis* in Lake Baikal (Kozhov, 1963; Kozhova and Ismest'eva, 1998), are most probably supported by convectional currents (Kelley, 1997; Granin et al., 1999). In winter, any nutrients discharged by rivers or released by decomposition in the water column or the bottom sediments accumulate under the ice and allow a more rapid growth of phytoplankton after ice

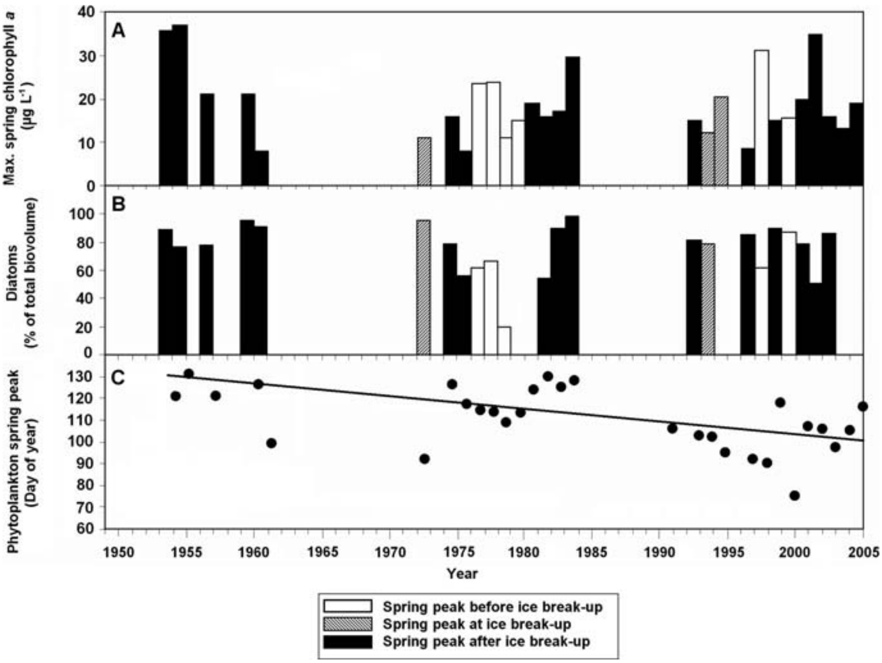

Fig. 14.3 Development of phytoplankton spring peak in Lake Erken based on at least weekly measurements in 1954–2005. **(A)** Maximum measured Chl a concentrations, **(B)** The percentage of diatoms in the total phytoplankton biovolume during spring peak. **(C)** Timing of the phytoplankton spring peak and its trend (Weyhenmeyer et al., 1999 updated with latest data).

break-up when light conditions improve. In deep lakes there may, however, be some delay due to complete vernal turnover as described in a later section.

In Western Europe, where the lakes are either ice-free or only freeze for a few days in the year (George, 2007), the most important winter effects are those associated with the year-to-year variations in the rainfall. In the English Lake District, George et al. (2004) showed that heavy winter rains tend to transport more dissolved reactive phosphate into the lakes but may also reduce the standing crop of phytoplankton by their flushing effect. The ecological response of the lakes to these flushing events is critically dependent on their residence time. In Blelham Tarn, a lake with an average residence time of 42 days, wet winters severely depleted the standing crop of phytoplankton (Fig. 14.4a). In contrast, in the North Basin of Windermere (Fig. 14.4b), a lake with an average residence time of 185 days, wet winters had no significant effect on the average biomass of phytoplankton. In some lakes, these flushing effects can even influence the composition of the phytoplankton much later in the year by reducing the size of the inocula that produce the early summer maxima.

In the deep perialpine lakes of Central Europe, the internal recycling of nutrients and the subsequent development of the phytoplankton are strongly influenced by the

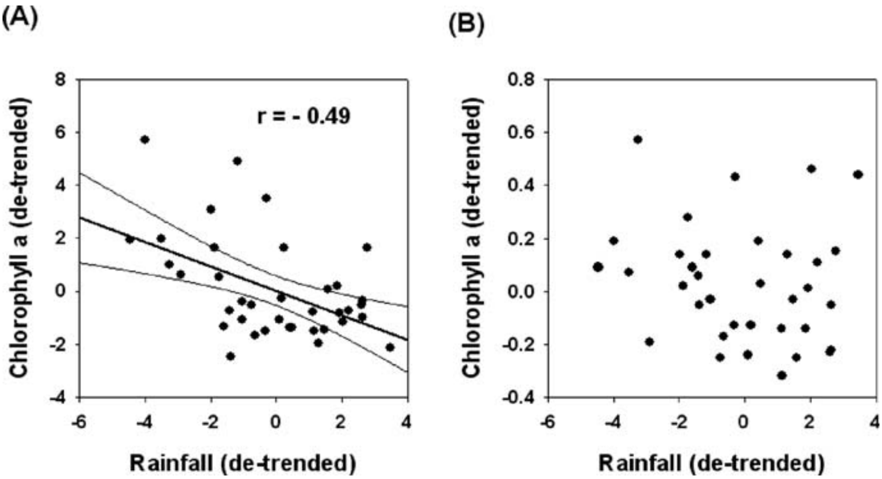

Fig. 14.4 Influence of winter rainfall on the de-trended winter concentration of phytoplankton chlorophyll in Blelham Tarn **(A)** and the north basin of Windermere **(B)**. (Data jointly managed by the FBA and CEH).

duration and intensity of vertical mixing in winter and early spring (Salmaso, 2002, 2005; Straile et al., 2003). The occurrence of several consecutive mild winters leads to incomplete mixing in such lakes, which further results in a gradual increase in deep-water temperature and a simultaneous decrease in deep-water oxygen concentrations. These gradual changes can be terminated by the occurrence of an unusually cold winter – or even an average winter, if the deep-water temperature has risen to a sufficiently high level. This then results in deep penetrative mixing, an abrupt fall in deep-water temperature and an abrupt rise in deep-water oxygen concentrations (Livingstone, 1997). Late winter and early spring may therefore be considered the most critical period in the annual cycle of deep lakes (Salmaso, 2005). During cold winters with a complete overturn in Lake Garda, total phosphorus concentrations in the euphotic layer exceeded those of milder winters by a factor of three and favoured the development of *Mougeotia* sp. and Oscillatoriales (Salmaso, 2002).

14.2.2 Spring

In lakes covered with ice, the disappearance of snow from the ice and the timing of break-up are crucial events for the development of the spring phytoplankton (Tulonen et al., 1994; Weyhenmeyer et al., 1999; Gerten and Adrian, 2000; Straile and Adrian, 2000). An earlier spring bloom in years following earlier ice break-up has been observed in lakes Müggelsee (Adrian et al., 1999) and Erken (Weyhenmeyer et al., 1999).

In Lake Erken, the spring peak of phytoplankton has advanced by about one month over the last 50 years (see Fig. 14.3C). Here, the timing of the ice-break and

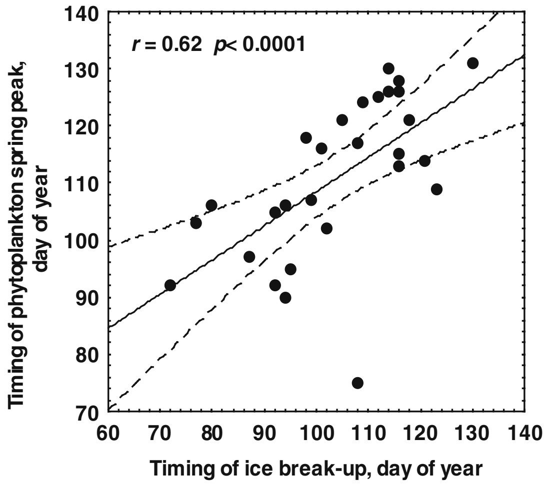

Fig. 14.5 Timing of the phytoplankton spring peak vs. the timing of ice breakup in Lake Erken for the period 1954–2005 (Weyhenmeyer et al., 1999 updated with more recent data)

the timing, composition and magnitude of the spring bloom (Fig. 14.5) depend on the weather experienced in March. In contrast, the duration of the spring bloom and the length of the post-bloom period are primarily controlled by nutrient availability (Blenckner, 2001). Since the size of the available nutrient pool differs from lake to lake, the decline of spring phytoplankton is not directly linked to the timing of ice break. Early ice-break combined with an early spring bloom may, however, result in an accelerated rate of nutrient depletion and an earlier decline in the early spring phytoplankton (Weyhenmeyer, 2001; Järvinen et al., 2006).

Owing to their smaller volumes, reduced heat storage, and shorter residence times, shallow lakes respond in a more direct way to inter-annual variations in the weather. In small, non-stratified, lakes, the climatic ‘signal’ captured during the spring turnover persists for only a short period of time (Gerten and Adrian, 2000). In contrast, some large but shallow lakes, like Lake Võrtsjärv, have an extended ‘climate memory’ and here the meteorological conditions experienced in winter and early spring determine to a large extent both the water level and the dynamics of the phytoplankton throughout the ice-free period (Nõges, 2004; Nõges et al., 2003).

In lakes without winter ice-cover, the timing of the spring bloom is not so strictly determined by one climatic variable. The transition period from winter to summer is smoother and the lakes ‘integrate’ the different elements of the climate signal in functionally different ways. In the lakes of the English Lake District, there is mounting evidence that the spring blooms are appearing earlier in the year. In the four Windermere lakes (Fig. 14.6) this trend is statistically significant only in two basins: the north basin of Windermere and Esthwaite Water. The observed rates of advance (4.2 and 5.8 days per decade) are at the upper end of that reported for the phenology of plants in terrestrial systems (Walther et al., 2002). The different responses observed in the individual lakes suggest that the timing of this key event is influenced by factors other than the water temperature. Initial analyses suggest

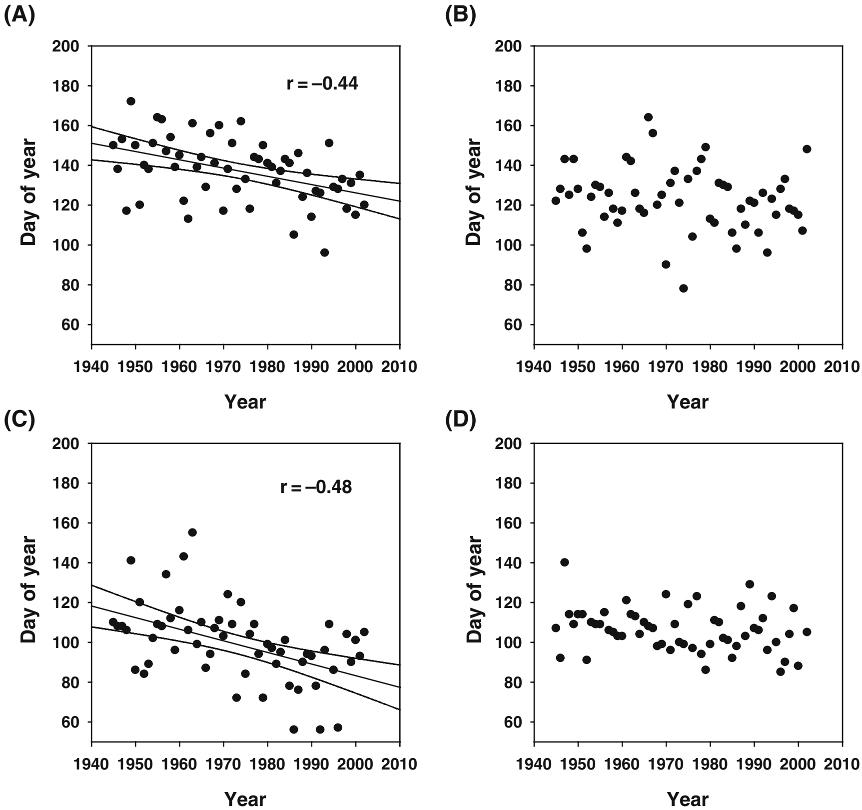

Fig. 14.6 The timing of the spring maximum of *Asterionella formosa* in four English lakes between 1945 and 2002. **(A)** North Basin of Windermere. **(B)** South Basin of Windermere. **(C)** Esthwaite Water. **(D)** Blelham Tarn. (Data jointly owned by the FBA and CEH – S.C. Maberly, unpublished).

that the timing of the bloom is controlled by a combination of factors that include warmer spring temperature, the increased availability of phosphate and the interannual variations in the rainfall (S.C. Maberly, unpublished). The apparent absence of a similar advance in Blelham Tarn may be attributed to the recently recorded increase in the winter rainfall (George et al., 2007) and the effect of the increased flushing rate on the development of the spring bloom.

In the deep lakes of Central Europe, spring mixing has a dual effect on phytoplankton development. Besides the replenishing effect of deep mixing on the nutrients in the epilimnion, the downward mixing of phytoplankton into aphotic layers seriously inhibits its growth (Huisman and Weissing, 1994; Steel and Duncan, 1999; Reynolds, 2006). In Lake Constance, a large and deep perialpine lake that seldom freezes, the onset of the spring phytoplankton bloom is largely controlled by turbulent diffusion, that is, by the transition from strong mixing in winter and early spring to weak mixing in summer (Peeters et al., 2007). Consequently, the onset of the bloom is closely correlated with the onset of thermal stratification, which in turn is determined by a complex interplay between temperature and wind (Peeters et al., 2007). Only when the depth of mixing is reduced to ca 40 m, does the improvement in the underwater light climate stimulate a marked increase in net growth rate of the phytoplankton which heralds the start of a bloom, (Peeters et al., 2007). A strong control of spring phytoplankton growth by vertical mixing seems to be rather a rule than an exception. As reported by Straile and Adrian (2000), phytoplankton growth in Lake Constance was inhibited by wind-induced reductions in the underwater light throughout the 16-year period (1979–1994) covered by their study.

14.2.3 Summer and Autumn

In many of the CLIME lakes, there has been a significant change in the phytoplankton growth patterns observed in early summer. At that time of year, many lakes experience a clear water phase, i.e. a period when the biomass of the phytoplankton declines sharply. In most cases, this decline has been related to zooplankton (*Daphnia*) grazing. Detailed accounts of this phenomenon have been given for lakes in Northern Europe (Gerten and Adrian, 2000; Weyhenmeyer, 2001; Adrian et al., 2006), Western Europe (Talling, 2003) and Central Europe (Straile, 2000; Straile and Adrian, 2000; Anneville et al., 2002a, b). In warmer years, a biomass of *Daphnia*, large enough to limit the growth of phytoplankton, is reached earlier in the year and results in an earlier and longer lasting clear-water phase. These effects appear to be lake specific: in shallow Müggelsee, spring water temperatures and *Daphnia* abundance both increased more rapidly than in large, deep Lake Constance. Consequently, the clear water phase started about three weeks earlier in Müggelsee than in Lake Constance (Straile and Adrian, 2000). The climatic responses observed in autotrophic species may not, however, be mirrored by heterotrophic species (Blenckner, 2005) since the processes responsible for the decay and recycling of the autotrophs are often lake specific. For example, in Lake Stechlin, spring diatoms simply sink as soon as the lake starts to stratify and the clear water phase is not connected with any grazing effects (Padisák, et al., 2003b). In some cases, the processes responsible for the breakdown and decomposition of the cells are markedly non-linear whilst in others they have critical thresholds. Cell death through parasitism may also account for significant proportion of phytoplankton loss in many lakes (Jassby and Goldman, 1974).

At some sites, causal links have been established between the meteorological conditions experienced in winter or early spring and events in the plankton the following summer. In large and shallow Lake Võrtsjärv in Estonia (270 km^2, mean depth 2.8 m), large year-to-year differences in the water level have a very pronounced effect on the development of the phytoplankton (Nõges, 2004; Nõges et al., 2003). The magnitude of the spring floods, determined largely by the winter air temperature and precipitation, explains most of the variability in annual mean water levels ($R^2 = 0.85$, $p < 0.0001$). When the level is low, the water is enriched with phosphorus by sediment resuspension and there is an associated reduction in the nitrate concentration due to denitrification. Since 1964, the phytoplankton biomass has been significantly lower in years of high water level, a pattern that was not related to any change in external loading of nutrients. These fluctuating water levels have also had an effect on the qualitative composition of the phytoplankton. During high-water periods, *Limnothrix redekei* and *L. planktonica* have typically accounted for more than 90% of the total wet weight of phytoplankton, which has remained under 30 g m^{-3}, even when the external nutrient loading was high. *Limnothrix* species can, by virtue of their shape and photoadaptive properties, maintain much higher growth rates than most other species when light levels are low (Gibson, 1987; Rücker et al., 1997). During low-water periods, the wet weight of phytoplankton has often exceeded 30 g m^{-3} and even reached a 100 g m^{-3} in the 1970s. During the 1990s and 2000s, low-water periods have also been characterized by an increase in the nitrogen-fixing species, *Aphanizomenon skujae*. Compared to *Limnothrix*, *Aphanizomenon* species need higher light levels (Foy et al., 1976) and are favoured by increased illumination in shallow water. Nitrogen fixation, recurrently measured in Lake Võrtsjärv in summers with low water levels, has likely been triggered by increased nitrogen losses due to denitrification (Tõnno and Nõges, 2003).

In Lake Geneva (Anneville et al., 2002a, b), the increase in the spring phytoplankton in warmer years, together with the decrease in phosphorus loading, have led to an earlier and more pronounced depletion of phosphorus in the productive layer. By mid-summer, the P-depleted layer often extends below the depth where the availability of light is the limiting factor (15–25 m). In these conditions, the mid-summer community is dominated by a complex of species, which are well adapted to low light levels and shorter days (e.g., *Mougeotia gracillima* and *Diatoma tenuis*). Species of this shape and size have an obvious competitive advantage when nutrients are low, due to their large surface to volume ratio and their size which renders them less vulnerable to grazing by zooplankton. Thus, these large species can survive and eventually achieve a greater biomass despite their relatively slow growth.

The most important weather-related effect observed in many of the CLIME lakes was the change in the timing of thermal stratification and the consequent extension of the summer growth period. These effects were particularly pronounced in Northern Europe where the extension of the growing season in several mesotrophic lakes mimicked the effects commonly associated with eutrophication. For example, in Lake Erken, Sweden, the oxygen concentrations recorded in the hypolimnion in late summer are critically dependent on the length of the stratification period and the transfer of heat into the deep water (Pettersson et al., 2003). A comparison of the oxygen concentrations measured in the hyplimnion in a series of relatively cold (1975–1979) and relatively warm (1994–2001) summers showed that the concentration in August decreased from an average of 4.2 mg l^{-1} to an average of 2.5 mg l^{-1} (data for 15 m depth differed significantly $p < 0.05$). This reduction was, in turn, responsible for a sustained increase in the concentrations of phosphate and ammonium in the hypolimnion and an intensification of internal nutrient cycling

(Chapter 15 this volume). A very similar deterioration in the quality of the water was observed in Heiligensee, a hypeutrophic lake in North Germany. Here, time series analyses for the period 1975–1992 showed an abrupt change in the structure of the phytoplankton community, centred on the late 1980s and early 1990s (Adrian et al., 1995). In Heiligensee, the earlier onset and longer duration of thermal stratification influenced the system in two different ways: (i) There was a rapid collapse of the spring diatoms as soluble reactive silicon was depleted and more cells were lost through sedimentation. (ii) There was a marked increase in the concentrations of phosphorus recorded in the hypolimnion at the end of summer that explained 69% of the variation in maximum algal standing stock recorded in the autumn.

In Western Europe, the growth patterns recorded during the summer have also changed in a systematic way. In the English Lake District, samples for chlorophyll analysis have been collected from four lakes at weekly or fortnightly intervals since 1964 (Talling, 1993). In these lakes, the growth cycle follows a ‘diacmic’ pattern with well defined maxima in the spring and summer. Despite this intensive sampling, the duration of the summer growth period can still only be estimated by fitting a Gaussian model to the raw observations. The model used has been described by George and Hurley (2004) and is based on fitting two Gaussian curves to the logarithm of the chlorophyll measurements:

$$ln(C(t)) = a + b_1 G(t; \mu_1, \sigma_1) + b_2 G(t; \mu_2, \sigma_2)$$

where t is the proportion of time that has elapsed from a defined starting date (31 December) and σ_2 is a measure of the duration of the summer growth period. Figure 14.7 a show the result of fitting this model to some example results for 1988. The samples were collected from the North Basin of Windermere, a relatively unproductive lake with a maximum depth of 60 m. Here, the fitted model explained a high proportion of the observed variation and provided a reliable measure of the ‘summer growth’ parameter ($\sigma 2$). Figure 14.7b shows the long-term change in the value of this parameter. In the 1970s and 1980s this parameter remained relatively constant, but it increased sharply in the 1990s. The Cumulative Sum (CuSum) plot in Fig. 14.7c shows that the pivotal year was 1993. An analysis of the internal and external factors responsible for this change showed that the key factor was the increased physical stability of the lake. Figure 14.7d is a CuSum plot of the change in the summer stability, as measured by the index described by Schmidt (1928). Statistical tests showed that the value of the $\sigma 2$ parameter only exceeded the ‘control limits’ in 1993, i.e. the year when there had been a marked increase in the summer stability of the lake. The phytoplankton species that dominate the open water in late summer are particularly responsive to the extension of the growing season. Most of these species grow quite slowly, so an additional cell division in late summer can have a major effect on their maximum biomass.

The other factor influencing the development of some slow growing species is the short-term variation in the intensity of wind-induced mixing. The lakes situated on the western seaboard of Europe are particularly sensitive to these physical effects (George, 2000a). In the more productive lakes in the English Lake District,

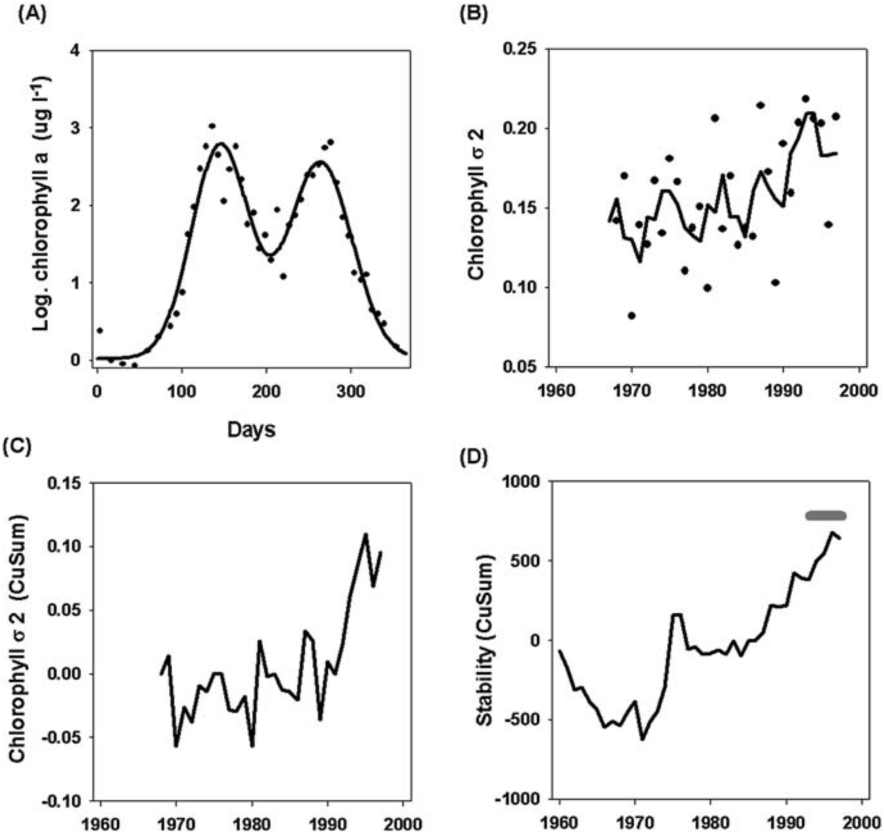

Fig. 14.7 Using a Gaussian model to quantify the change in the duration of the summer growth period in the North Basin of Windermere. A – an example fit for 1988; B – the long-term change in the σ2 parameter, an empirical measure of the summer growth period; C – The Cumulative Sum (CuSum) plot for the σ2 parameter; D – The CuSum plot of the long-term change in the summer stability of the water column (measured by the index described by Schmidt (1928)). (Data jointly managed by the FBA and CEH).

the seasonal succession of phytoplankton follows a remarkably predictable pattern (Reynolds, 2006). The first species to dominate are mostly diatoms which are then replaced by motile flagellates. These small forms have high rates of growth but can only survive if the water column is periodically mixed by the wind. When the water is warm and there is relatively little wind, slow growing forms like the bloom forming species of cyanobacteria become dominant. Figure 14.8 shows the effect that year-to-year changes in the weather had on the growth of the blue-green alga *Aphanizomenon* in Esthwaite Water between 1956 and 1972. Once the time-series had been de-trended to remove the effects associated with enrichment, there was a striking correlation ($r = 0.82$, $p < 0.001$) between the summer abundance of the *Aphanizomenon* and the stability of the lake. The factors influencing the growth of

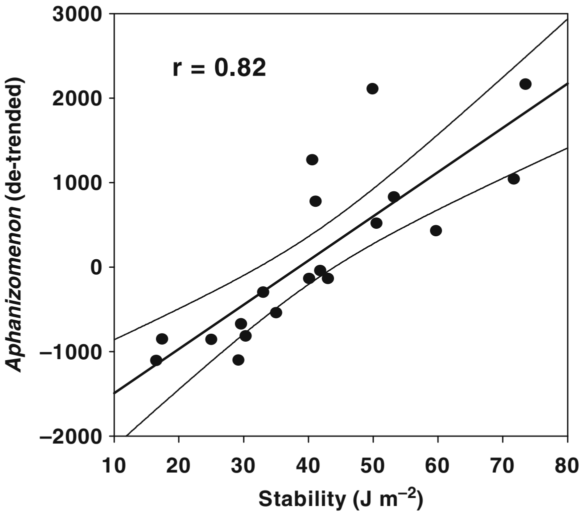

Fig. 14.8 The relationship between the abundance of Aphanizomenon in Esthwaite Water and the stability of the water column in summer. The Aphanizomenon time-series was de-trended to minimise the effects of progressive enrichment. The stability was calculated using the procedure described by Schmidt (1928)

Aphanizomenon in Esthwaite Water have been discussed in more detail by George et al. (1990) and George (2000b). The pattern described is quasi-cyclical and appears to be governed by the position of the Gulf Stream in the eastern Atlantic.

The results of a recent mixing experiment in Lake Nieuwe Meer, a hypertrophic lake in the Netherlands, demonstrates that quite short periods of reduced mixing can have a major effect on the growth of bloom-forming cyanobacteria (Jöhnk et al., 2008). This experiment was conducted during the hot summer of 2003 and showed that the summer heatwave was the main factor responsible for a dense *Microcystis* bloom. Simulations with a coupled hydrodynamic-phytoplankton competition model showed that high temperatures favoured cyanobacteria both directly, through increased growth rates, and indirectly, by reduced turbulent mixing.. The European summer heatwave of 2003 is considered by many to be a prototype of future summers in the region. Results of this kind imply that, whilst high air temperature *per se* has an effect on the growth of these species, it is the combination of high temperatures and reduced wind-speed that provides the conditions necessary for the appearance of dense surface blooms.

The growth and development of the autumn phytoplankton is also critically dependent on the mixing characteristics of the lakes. In stratified lakes, the high nutrient concentrations that accumulate in deep water during extended stagnation periods (Adrian et al., 1995) can promote strong water blooms when the overturn starts (Pettersson et al., 2003; Kangro et al., 2005). In polymictic lakes, such as Peipsi and Võrtsjärv, phytoplankton biomass usually increases during the autumn and only declines after the lakes start to freeze (Nõges et al., 2004). When the ice-cover is late, the biomass peak appears later and, since the cells sink slowly in cold water, higher than average biomasses may be recorded for as long as two months after the lakes have frozen. Very deep lakes may have even longer lasting 'memories' of antecedent conditions and extreme climatic events (Chapter 17 this volume). There is even evidence to suggests that the meteorological conditions experienced during the winter can affects the phytoplankton in the following summer and autumn but the mechanisms are complex and not yet clear (Straile et al., 2003).

14.3 The Impact of Climate Change on the Structure of the Phytoplankton Community

14.3.1 Phytoplankton Species Favoured by Climate Change

Climate can be considered the major factor determining the distribution of species at a continental scale (Pearson and Dawson, 2003). Small variations in climate can have dramatic effects on biota, especially in extreme habitats, where many species live at the limit of their environmental tolerances. Distribution, composition and species diversity of diatoms in Sub-arctic Lapland are strongly regulated by temperature and other climate-related factors (Weckström and Korhola, 2001; Sorvari et al., 2002). Global warming is projected to cause a northward extension of those species that are better adapted to higher temperatures. As an example, the bloom-forming cyanobacterium *Cylindrospermopsis raciborskii* is causing increasing concern because of its potential toxicity and invasive behavior at middle latitudes (Padisák, 1997; Briand et al., 2004; Paerl and Huisman, 2008). *C. raciborskii*, originally classified as a tropical to subtropical species with a higher temperature optimum than most cyanobacteria (Gorzó, 1987), has now been reported in several Central European countries, such as Hungary (Tóth and Padisák, 1986), Austria (Dokulil and Mayer, 1996), France (Couté et al. 1997), Germany (Krienitz and Hegewald, 1996; Stüken et al., 2006) and Poland (Stefaniak and Kokocynski, 2005). In Lake Balaton *C. raciborskii* blooms now appear in years when summer temperatures are significantly higher than average and there is an active P-pool in the sediments (Padisák, 1998). In the consecutive dry years 2000–2003, the temperature requirements of the species were fulfilled but the sedimentary P-pool must have been insufficient since no increased growth was actually observed (Padisák et al., 2006).

The increased incidence of metalimnetic or upper hypolimnetic maxima of the cyanobacterium *Planktothrix rubescens* (considered conspecific with *P. agardhii* by Humbert and Le Berre, 2001) is another phenomenon reported from many stratified lakes (Dokulil and Teubner, 2000; Davis et al., 2003; Teubner et al., 2003, 2006; Padisák et al., 2003a; Anneville et al., 2004; Jacquet et al., 2005; Salmaso, 2005). This species is particularly efficient at harvesting light due to its high phycobiliprotein content (Bright and Walsby, 2000; Greisberger and Teubner, 2007) and ability to optimize its position in the water column (Reynolds et al., 1987). Walsby and his coauthors (Walsby, 2005; Walsby et al., 2006) used a modelling approach to demonstrate that the ability of the *Planktothrix rubescens* to stratify in Lake Zürich was related to the size and shape of its filaments, which respond to the irradiance by changing their density. This model was also used to explain the Burgundy-blood phenomenon sometimes observed in Lake Zürich in November and December when, after deeper mixing and lower insolation, *Planktothrix* filaments become buoyant and float to the surface in subsequent calm periods.

Since the 1950s, the eutrophication of many deep, alpine lakes has led to the progressive suppression of *Planktothrix*, as the light reaching the metalimnion became insufficient (Sas, 1989). The recent success of *P. rubescens* in a number of lakes is most probably caused by a synergetic effect of increased transparency due to the reduction in the phosphorus loads, the deepening of the P-depleted zone and increased water column stability (Anneville et al., 2005; Jacquet et al., 2005; Teubner et al., 2003, 2006). The physiological shift from autotrophic to photoheterotrophic metabolism seems to be crucial for the success of the *P. rubescens* layers that develop below the compensation point. There is also experimental evidence to suggest that acclimatisation to dim-light stimulates the uptake of organic compounds by *P. rubescens* (Zotina et al., 2003).

The depth of winter mixing affects the development of *P. rubescens* in a nonlinear way. Bürgi and Stadelmann (2002) suggested that deeper mixing in Lake Sempach (maximum depth 87 m) enhanced the competitiveness of the species by extending the low irradiance zone (see also Bright and Walsby, 2000; Greisberger and Teubner, 2007). However, if the mixing depth exceeds the mean critical pressure depth around 90 m, the gas vesicles in this species collapse (Walsby et al., 1998; Bossard et al., 2001). That observation supports the suggestion made by Anneville et al. (2004) that reduced winter mixing (to 60–100 m) in Lower Zürich lake (maximum depth 136 m) in the 1990s may have contributed to the increased abundance of *P. rubescens* during winter. These examples suggest that phytoplankton composition is more sensitive to climate change than is overall phytoplankton biomass. As a result, year-to-year variation in weather may cause synchronous changes in phytoplankton composition over a wide geographical area but smaller effects on biomass.

In some of lakes studied in Central and Northern Europe, recent reductions in their nutrient loads and their subsequent re-oligotrophication further complicates the analysis of change. Jankowski et al. (personal communication) analysed the simultaneous effects of re-oligotrophication and climate variability on phytoplankton diversity using data from CLIME lakes in Central and Northern Europe. They found that lake restoration, i.e. a reduction in the phosphorous loads, has resulted in an increase in phytoplankton diversity. Over the last 25 years, the number of genera reported from these lakes has increased by 20–70%. This trend appears to be universal and was not related to the trophic status of the lakes.

14.4 Discussion

Our study on the response of phytoplankton to climatic change across lakes in Europe has shown that systematic changes in the weather have already had a significant effect on the seasonal dynamics of phytoplankton at a number of CLIME sites.

Many of these changes can be directly related to observed large-scale changes in the climate and regional variations in the circulation of the atmosphere. In winter, the most important effects were those associated with the inter-annual variations in the North Atlantic Oscillation (Chapter 17, this volume), i.e. the duration of ice-cover, rainfall, and wind-induced mixing. In summer, these effects were less pronounced but the dynamics of the thermally stratified lakes was then influenced by the year-to-year variations in the number of warm days with very little wind (Chapter 16 this volume) that had a strong effect on bloom forming cyanobacteria.

In the late 1980s, the atmospheric pressure gradient quantified by the NAO winter index changed in a systematic way with the index remaining in its positive (mild winter) phase for several years in succession. This shift was accompanied by a major change in the lake temperature regimes observed over most of Europe (Chapter 6 this volume). This effect was most pronounced in 1988 and was manifested as an upward jump separating the earlier 'cold' period from the following 'warm' period (Adrian et al., 1995; Weyhenmeyer, 2001; Weyhenmeyer et al., 2002; Anneville et al., 2005). In some cases, the effects of these mild winters could be detected much later in the year, e.g. when the earlier ice break-up and warming led to an earlier onset of stratification and the growth of summer phytoplankton. In the large lakes of Sweden, for example, the 'shift' recorded in the early 1990s resulted in an extension of the growing season by at least one month (Weyhenmeyer, 2001). Different phytoplankton groups responded differently to this sudden warming. Although there was no increase in the total biomass recorded between May and October, the biomass of temperature-sensitive groups, such as the cyanobacteria and chlorophytes, increased in spring and early summer. Very similar patterns have been observed in a series of 17 lakes studied in Switzerland, Germany, Sweden and the UK. When the composition of the phytoplankton was compared in two contrasting years (1987 and 1989), much higher biomasses were recorded during the warm winter of 1989 (Weyhenmeyer et al., 2002). Cyanobacteria were most affected, and their annual mean biomass increased by a factor of as much as 100 when the two extreme years were compared. Several recent studies (Paerl and Huisman, 2008, 2009; Jöhnk et al., 2008) have shown that climate change is a potential catalyst for the further expansion of harmful cyanobacteria in eutrophic lakes. Rising temperatures, reduced cloud cover in combination with high nutrient loading all favor cyanobacterial dominance. Blooms increase the turbidity of lakes, can deplete oxygen levels, and often produce a bad smell. Moreover, many species of cyanobacteria can produce toxins that can cause serious liver, digestive, neurological, and skin diseases in animals as well as humans. More studies are needed on the factors influencing the development of toxic versus non-toxic strains of cyanobacteria. It has recently been shown that competition between toxic and non-toxic strains of *Microcystis* is strongly influenced by the light regime experienced in the critical spring-summer period (Kardinaal et al., 2007).

The effects associated with the inter-annual variation in the summer weather were most pronounced in the more productive lakes of the English Lake District. Here, short-term changes in the stability of the water column had a major effect on the seasonal development of bloom-forming species of cyanobacteria, such as *Anabaena*, *Aphanizomenon*, and *Microcystis*. In the last forty years, there has been a significant increase in the number of calm, anticyclonic days recorded in this region during the summer (Briffa et al., 1990). George (2006) has shown that these changes have had a direct effect on the physical stability of the lakes and an indirect effect on the seasonal development of the plankton (George and Taylor, 1995). Such 'extreme events' have also had some effect on the seasonal development of the phytoplankton in larger, less productive lakes. For example, in 2002, the warmest summer ever recorded in northern Europe, the increased thermal stability of Lake Mälaren in Sweden resulted in a greatly increased consumption of oxygen in deep water, much higher concentrations of nutrients in the hypolimnion and nutrient depletion in the surface water. In autumn, the sudden transfer of nutrients from deep water, combined with the high water temperatures, resulted in an unusually intense bloom of cyanobacteria. The severity of these blooms may also have been influenced by the rainy periods experienced the preceding year. These periods led to a distinct increase in the chemical loading of Lake Mälaren and an associated increase in the colour of the water (Weyhenmeyer et al., 2004). Significant increases in lake water colour and DOC concentrations have recently been reported from a number of European lakes (see Chapter 12 this volume). Such changes are known to have a positive feedback effect on lake surface temperatures and can lead to the development of steeper and longer lasting periods of thermal stratification as more heat is absorbed near the surface. High concentrations of phytoplankton also act as an optically active substance in the water, absorbing and scattering the downwelling irradiance and storing more heat in the upper part of the water column (Arst, 2003; Paerl and Huisman, 2008). The surface temperature within cyanobacterial blooms in Lake IJsselmeer, Netherlands, was 3°C above ambient waters (Ibelings et al., 2003). This could represent an important positive feedback mechanism, whereby buoyant cyanobacteria locally enhance surface temperatures, which in turn favors their competitive dominance over eukaryotic phytoplankton (Hense, 2007). Summer heat-budget calculations for two large enclosures installed in Blelham Tarn (English Lake District) demonstrated convincingly that more heat was absorbed near the surface and more lost by night-time cooling in the enclosure when the concentration of phytoplankton was very high (Jones et al., 2005). In this enclosure, the increased absorption of solar radiation at the surface and the decreased penetration of light both raised the thermocline and strengthened the temperature gradient. The authors suggested that these changes would have quite a complex effect on the algae by increasing their growth rates in the mixed layer whilst reducing the overall depth of the euphotic zone. The importance of water clarity relative to wind mixing in determining the mixing depth or the depth of thermocline or mixing decreases with increasing lake size and has little effect on lakes with a surface area greater than 5 km^2 (Fee et al., 1996). As the world becomes warmer, the consequent intensification of stratification is likely to increase the depletion of oxygen in deep water, inhibit the transfer of nutrients from the hypolimnion and limit the vertical movement of passively floating algae.

Analyses of the kind reported here, where changes in the composition and seasonal dynamics of lake phytoplankton are related to long-term changes in the climate have inherent strengths and weaknesses. In some cases, the processes responsible for this linkage are quite clear but in others this is still a matter for some speculation. These observations are, however, very useful for validating the results of simulation models (Chapter 15 this volume) and quantifying the impact of extreme climatic events. In many CLIME lakes, the observed changes in the phytoplankton could be related unequivocally to changes in the physical characteristics of the lakes. The main weakness of the observational approach is that there are several interfering processes such as eutrophication, reoligotrophication, changes in acid deposition and so on, that go on in parallel with the climatic variations and influence the growth of the phytoplankton in similar ways. The sensitivity of phytoplankton communities to climatic signals is further complicated by issues such as geographic location, lake type and the trophic status of the individual lakes. In this respect, the spatial coherence analyses described in Chapter 17 can prove most illuminating especially if combined with the functional group approach advocated by e.g. Reynolds et al. (2002). The studies carried out on a homogenous set of peri-alpine lakes (Anneville et al., 2004, 2005) represent a good example of this approach but the data sets acquired from other regions are still underexploited.

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Chapter 15
Modeling the Effects of Climate Change on the Seasonal Dynamics of Phytoplankton

Thorsten Blenckner, Alex Elliott, Hampus Markensten, Charlotta Pers, and Stephen Thackeray

15.1 Introduction

Changes in the climate, on the scale now predicted, will have a profound effect on the biomass and species composition of lake phytoplankton (Adrian et al., 1995; Weyhenmeyer et al., 1999; Winder and Schindler, 2004; De Senerpont Domis et al., 2007). Phytoplankton species have short life-cycles and their growth is influenced by a variety of chemical, biological and physical factors (Reynolds, 1984). Year-to-year variations in the weather are known to have a major effect on the seasonal dynamics and growth of phytoplankton (Chapter 14, this volume). For example, short-term changes in the mixing characteristics of a lake regulate both the growth and spatial distribution of cyanobacteria (Soranno, 1997; Jöhnk et al., 2008). There are, however, large between-lake variations in these climatic responses that are principally related to the sensitivity of the sites (Magnuson et al., 1990; Blenckner, 2005; George, 2006). Most climatic effects are mediated by hydrodynamic factors such as, the timing and intensity of thermal stratification and variations in the water level. One approach to the quantification of these effects is the analyses of long-term data acquired by intensive (i.e. weekly or bi-weekly) sampling. Statistical analysis of this kind cannot, however, unequivocally identify the most important driving processes or produce projections for future 'warm world' conditions. To explore these effects, we either need to design expensive field experiments or simulate the seasonal dynamics of the lakes by numerical modelling. In CLIME, we used a combination of numerical models to simulate the physical and chemical responses of the lakes and the dynamics of phytoplankton growth. Details of the model used to simulate the physical responses of the lakes are given in Chapter 7, this volume whilst Chapters 9, 11 and 13 summarise some results from the catchment-based models.

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In the last decades, many different mathematical models have been developed to explore the dynamics of natural ecosystems and their response to change (see for example Jørgensen, 1995). These models are a representation of the understanding gained from our experience, observations, measurements and experiments and formalize that understanding in a conceptual model or a set of equations. It is, however, important to recognize that, the output of these models is not reality but a summary of possible outcomes which can still be very helpful for decision making.

In general, two different approaches have been used to model aquatic ecosystems. The first is the minimal approach (e.g. Huisman et al., 2002; Huppert et al., 2002) where the model simplifies and minimises the key processes and produces outputs that cannot readily be compared with field data. These models nevertheless can provide useful insights into ecological mechanisms (Peeters et al., 2007) as long as their outputs are carefully scrutinized (van Nes and Scheffer, 2004). The second is the complex ecosystem model, which attempts to simulate the processes controlling phytoplankton dynamics under natural conditions (e.g. Reynolds and Irish, 1997). These models are better at representing natural events, but typically have a complex structure and a high number of defining parameters. The interactions between the different model parameters can also be complex and can obscure key processes by subtle, internal compensations. However, this approach is a first step to model the complex nature of ecosystems and only by this step we can improve the parameterization of specific processes and advance our understanding.

In this chapter, we explain how we used a combination of lake and catchment models to quantify the climatic responses of a number of lakes located in Northern and Western Europe. The intention here is not to reproduce the more detailed descriptions and results presented elsewhere (e.g. Reynolds et al., 2001; Pers, 2002; Schneiderman et al., 2002; Persson et al., 2005; Markensten and Pierson, 2007), but to present a series of Case Studies that show how these models can be combined to address a range of climate-related questions.

One of the major challenges of using models that are driven by the outputs from other models is that the uncertainty of the output is increased by the cumulative uncertainties of the models in the cascade (Beven, 2006). This is particularly true for approaches using climate projections, which have high levels of natural variability. In the examples presented here, we start with a simple model that simulates the monthly variation in the supply of phosphorus and end with a more complex model that predicts the day-to-day variations in the growth of phytoplankton (see Fig. 15.1). In all the examples, the meteorological driving variables were those produced by the Regional Climate Models (RCM) described in Chapter 2 (see also Döscher et al., 2002; Jones et al., 2004; Räisänen et al., 2004). The boundary conditions for these RCMs are those generated by two contrasting general circulation models: the HadAM3 GCM (Hadley Centre, UK Met Office, Gordon et al., 2000) and the ECHAM4/OPYC3 GCM (Max Planck Institute for Meteorology, Germany, Roeckner et al., 1999).

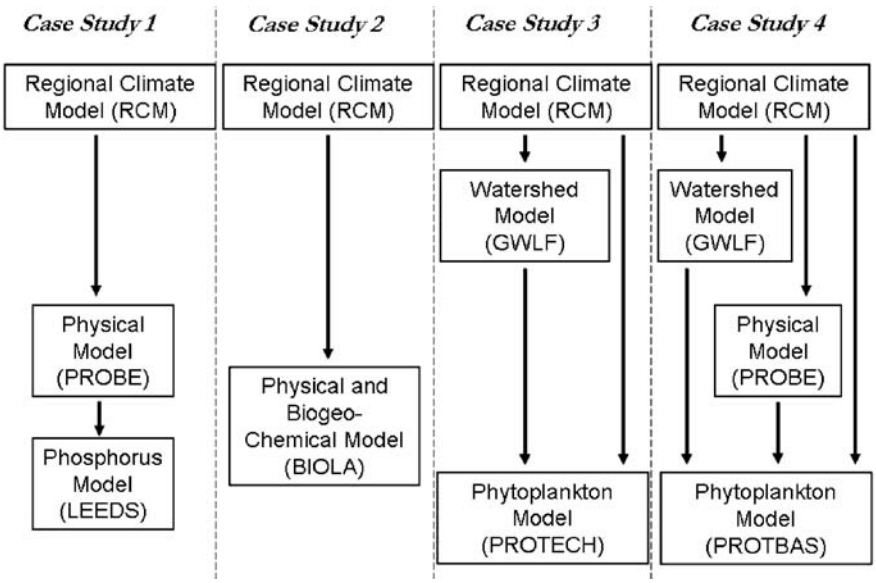

Fig. 15.1 The model combinations used in the Case Studies described in this chapter

15.2 Models Used in the Case Studies

15.2.1 Regional Climate Model

In all the Case Studies, the models were driven by the downscaled outputs from the regional climate models collated by the Swedish Meteorological and Hydrological Institute (SMHI) described in Chapter 2. These RCMs had a horizontal resolution of 0.44° (approximately 49 km) and were perturbed by the A2 and B2 emission scenarios described by the IPCC (2000). For Case Studies 1, 3 and 4 we used the results generated by the 'fixed period' version (hereafter called pRCM) with a control run that covered the period between 1961 and 1990 and scenario runs for the period between 2071 and 2100. For Case Study 2, we used a 'transient' version of RCM (hereafter called tRCM) driven by the ECHAM4/OPYC3 GCM and the B2 emission scenario running continuously from 1961 to 2100.

15.2.2 Physical Lake Model

The PROgramme for Boundary layers in the Environment (PROBE) model is an equation solver that has been applied to a wide range of environmental studies (Svensson, 1978; Sahlberg, 1988; Ljungemyr et al., 1996). PROBE produces a 1-dimensional description of temperature variations and heat fluxes in the water column and the atmosphere that serves as the upper boundary. The geographical location and depth-area curve for the target lake has to be specified before the model can be perturbed by an appropriate combination of air temperature, wind speed, humidity and cloud cover values. Further details of the PROBE model are given in Chapter 7, this volume. In the version used here, the structure and dynamics of the water column were simulated at 1 m intervals and the model run with a temporal resolution of 6 hours. The model was calibrated with data acquired from a local meteorological station and an automatic station that provided hourly measurements of the water temperatures at three different depths (0.5, 3, 14 m). The model was driven with the output from RCM for the control period and validated with water temperature measurements from the same periods. Thereafter, it was perturbed by the outputs from the RCM driven by the A2 and B2 emission scenarios.

15.2.3 Phosphorus Model

The Lake Eutrophication, Effect, Dose, Sensitivity model (LEEDS) is a dynamic model that produces monthly predictions of the phosphorus fractions in nine different lake compartments. It includes compartments for dissolved, colloidal and particulate phosphorus in both surface and deep water, phosphorus in ET-sediments (i.e. the areas of sediment exposed to erosion and transport), A-sediments (i.e. the areas of fine sediment accumulation) and the standing crop of phytoplankton. The phosphorus pathways include the major inflows and outflows, sedimentation, burial in A-sediments, resuspension from ET-sediments, release from A-sediments, mineralization, vertical mixing, bio-uptake and turnover (i.e. transformation from phytoplankton to particulate phosphorus). The required input data are geographic position, catchment area, lake area, mean lake depth, mean monthly wind speed, mean annual precipitation and the phosphorus concentration in the inflows. The meteorological data used to drive the model were the records from Uppsala airport and the projections from the RCM simulations. The model was calibrated with weekly (Erken) or monthly (Ekoln and Galten) total phosphorus and phosphate data acquired from the epilimnion and hypolimnion between 1994 and 2000. A more detailed description of the model can be found in Malmaeus and Håkanson (2004).

15.2.4 Biogeochemistry Model

The BIOgeochemical LAke model, BIOLA, uses the PROBE model to estimate the daily and vertical variations in the water temperature and a process-based biogeochemical component to simulate the associated variations in a number of chemical and biological variables (Pers, 2002). The structure of the model is shown in Fig. 15.2. It is based on 14 state variables that include the concentration of oxygen, inorganic nutrients and organic matter in the water and in the sediment and a limited number of biological variables. The parameters in

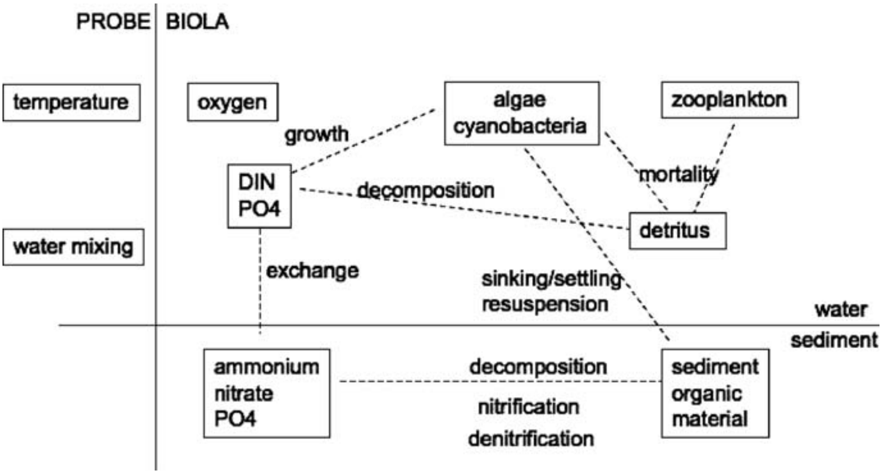

Fig. 15.2 Schematic view of the BIOLA model. *Broken lines* are fluxes and *boxes* show the effects of temperature, wind mixing and biogeochemical fluxes. DIN = Dissolved inorganic nitrogen, PO4 = phosphate

the model are based on established estimates of the rate processes, which transforms carbon, phosphorus and nitrogen in lakes. The model simulates the seasonal variation in nutrient concentration and a range of biological variables. Important processes included in the model are; phytoplankton growth, zooplankton grazing, decomposition of organic material, exchange of nutrients with the sediments and de-nitrification.

The model was calibrated using observed meteorological data (1994–1999) and perturbed with the six hourly outputs generated by the transient RCM (tRCM). The critical variables were: the geographical location, the depth-area distribution of the lake, wind speed, short-wave and long-wave radiation, relative humidity and the air temperature. When preliminary tests indicated too strong mixing, the wind forcing was reduced by 30% in subsequent simulations. Calibration of the biogeochemical variables was difficult due to the number of inter-dependencies. Model parameters were thus adjusted for the phytoplankton, cyanobacteria, dissolved inorganic nitrogen (DIN), phosphate, and oxygen to produce the best fit for the weekly observations. Estimates were produced for the concentrations of nutrients in the epi- and hypolimnion, the vertical distribution of oxygen and the concentration of phytoplankton and cyanobacteria in the epilimnion. All variables were compared with observations and the parameters adjusted with a ‘sensitivity analysis’ where one parameter was changed at a time. The physical part of the model was calibrated using manual temperature-depth profiles acquired by weekly routine sampling between 1990 and 2000 and validated against a period with unusually warm weather (2000–2002) based again on weekly measured temperature-depth profiles.

15.2.5 Catchment Model

The catchment model used is the Generalized Watershed Loading Functions (GWLF) a formulation which simulates monthly dissolved and total phosphorus and nitrogen loads in catchments with mixed land uses (Haith and Tubbs, 1981; Haith and Shoemaker, 1987). Further details of the GWLF model are given in Chapter 3 and more example applications in Chapters 9, 11 and 13. GWLF is a hybrid model that combines a dynamic representation of the catchment hydrology with a simplified ‘export coefficient’ approach to the flux of nutrients. The model is driven by daily temperature and precipitation data, water balances are calculated at daily intervals and streamflow includes contributions from surface runoff and groundwater discharge. Runoff is calculated using the SCS curve number method (Ogrosky and Mockus, 1964). In the example presented here, curve numbers were estimated from the soil and land-use data available for the Mälaren catchment. Dissolved nutrient loads are derived by multiplying runoff by a land-use-specific nutrient concentration. The effects of particular agricultural practices, such as manure spreading, can also be included if this information is available. The hydrological component of the model was calibrated using daily meteorological data acquired from a 4 × 4 km grid interpolation of local recording stations (Johansson, 2004). The calibration covered the period between 1980 and 1991 and the model fit optimised using measured stream discharge. Simulations for the historical period were based on a 30-year period (1961–1990). In the ‘warm world’ simulations it was assumed that land use and management remained constant.

15.2.6 Phytoplankton Model

15.2.6.1 PROTECH

PROTECH (Reynolds et al., 2001) is a model that simulates the growth and succession of different functional groups of phytoplankton. The biological ‘core’ of the model are the equations that define the daily change in the chlorophyll *a* concentration (X) of each algal species:

[1] $\Delta X/\Delta t = (r' - S - G - D)X \qquad \text{mg m}^{-3}\ \text{d}^{-1}$

where r' is the growth rate defined as a proportional increase over one day, S is the loss from settling in the water column, G is the loss from grazing (species > 50 μm are assumed not to be grazed) and D is the loss from dilution. The growth rate (r') is further defined by:

[2] $r' = \min\{r'_{(\theta,I)}, r'_{\mathrm{P}}, r'_{\mathrm{N}}, r'_{\mathrm{Si}}\} \qquad \text{d}^{-1}$

where $r'_{(\theta,I)}$ is the growth rate related to temperature and daily photoperiod and r'_{P}, r'_{N}, r'_{Si} are the growth rates determined by phosphorus, nitrogen and silica concentrations.

PROTECH includes its own hydrodynamics sub-routine, where the mixing depth is calculated for each daily time-step using the Monin-Obukhov equation and estimates of the heat fluxes and wind stress (Imberger and Hamblin, 1982). The model also calculates the changes in nutrient concentrations (soluble reactive phosphorus, SRP; PO_4-P), nitrate-nitrogen (NO_3-N) and silica (SiO_2) with respect to inflow concentrations, losses via the outflow and biological uptake. In the example presented here, the model was calibrated with daily measurements of the wind speed, daily cloud cover, inflow volume and associated nutrient concentrations for the period between 1997–1999. Additional nutrient inputs from a fish farm and a sewage treatment works were also included and the model perturbed with the downscaled outputs from the HadAM3 regional climate model (pRCM) developed by the UK Meteorological Office using the A2 and B2 scenarios produced by the IPCC (IPCC, 2000).

15.2.6.2 PROTBAS

The PROTBAS (PROtech Based Algal Simulations) model is based on PROTECH and uses the same input data to calculates the daily increase in the biomass of different species of phytoplankton, measured as chlorophyll *a*. The procedure used for calculating the temperature and light-corrected growth rate of the phytoplankton is, however, different to that used in PROTECH (Markensten, 2005). In PROTBAS, light limited growth is assumed to occur at light levels below I_k, where I_k is derived from the slope of the light limited portion of the light versus growth relationship for each functional group. It assumes that at light intensities below I_k, phytoplankton growth rate is limited by light and not primarily by temperature. When the phytoplankton cells are not light saturated, their pigment content, pigment packaging size and shape can change and override any effects associated with any temperature-dependent enzyme activities (Markensten and Pierson, 2007).

15.3 The CLIME Case Studies

15.3.1 Case Study 1: The Internal Recycling of Phosphorus in Three Swedish Sites

In this Case Study, we used the outputs from the regional climate model (pRCM) and the physical lake model (PROBE) to drive the dynamic phosphorus model (LEEDS) to simulate the effects of changes in the weather on the internal cycling of phosphorus in three Swedish lakes (Table 15.1). The lakes are all located in central Sweden but are morphometrically very different and have residence times that range from one month to seven years (Malmaeus et al., 2005).

In this simulation, the PROBE output was used as the input to the LEEDS model, which then produced estimates of the assimilation and recycling of phosphorus in

Table 15.1 The characteristics of the three Swedish sites used in Case Study 1

Lake	Mean depth (m)	Max. depth (m)	Residence time (year)	Trophic status	Case study
Erken, Sweden	9	21	7	Mesotrophic	1, 2
Galten, Mälaren, Sweden	3.4	19	0.07	Hypereutrophic	1, 4
Ekoln, Mälaren, Sweden	11.5	50	1.2	Eutrophic	1

the three lakes. The monthly discharge and phosphorus loads used in the simulations were the averages calculated for the last ten years (Malmaeus et al., 2005). In all cases, it was assumed that the external loading remained the same since the main aim of the study was to compare the climatic sensitivity of the individual lakes. The control and future simulations in Fig. 15.3 are a statistical summary of the changes projected between 2071 and 2100, not a sequential representation of the changes expected in a particular year, i.e. the first year shown on the X-axis is not necessarily 2071.

The results show that the projected changes in the climate have a very different effect on the internal recycling of phosphorus in the three lakes. In the two lakes with the shorter residence time (Galten and Ekoln), there was no significant change in the internal flux of phosphorus under 'warm world' conditions. The average annual concentration of total phosphorus projected for the lakes for 2071–2100 was also very similar to that recorded between 1961 and 1990. In contrast, in the lake with the longest residence time (Erken), there was a marked intensification of the phosphorus cycle as more phosphate was released from the sediment during the extended period of thermal stratification. Here, the average concentration of total phosphorus in the epilimnion almost doubled under the conditions projected by the 'warm world' A2 emission scenario. Short-lived increases in the amount of phosphorus released by the sediment were also predicted for Ekoln during periods of strong stratification but the relatively short residence time of that lake ensured that these episodes had little effect on the mean phosphorus concentration. Taken together, these results imply that the internal recycling of phosphorus will increase as the world becomes warmer. The magnitude of this response will, however, vary from lake to lake and may be masked in lakes with very long hydrological residence times.

In addition to the residence time, the projected change in the stability of Lake Erken also lead to a higher mineralization rate, elevated bacterial activity, a decreased concentration of oxygen in and higher rates of phosphate diffusion from the sediment into the water column. As a consequence, the biomass of phytoplankton in the lake increases and leads to a further intensification of the phosphorus cycle.

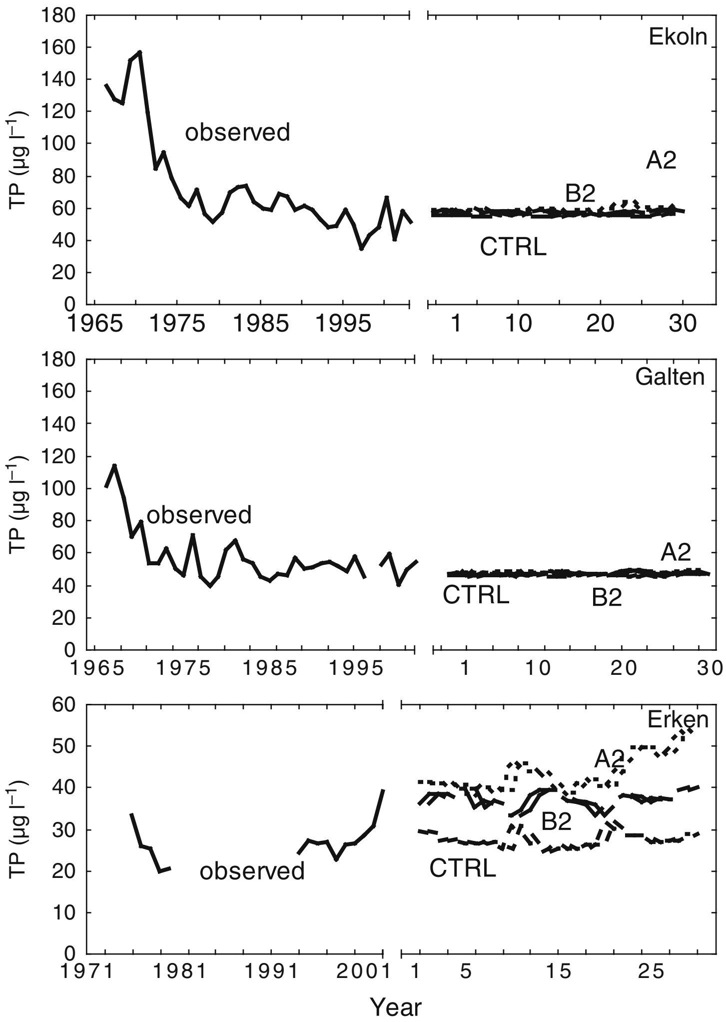

Fig. 15.3 The historical and simulated variation in the total phosphorus (TP) concentration ($\mu g\ l^{-1}$) in the three Swedish lakes (Ekoln and Galten and Erken). (CTRL = 1961–1990, two future emission scenarios, A2 and B2 = 2071–2100. Note the different scales on the Y-axis

15.3.2 Case Study 2: The Long-Term Change in the Physical, Chemical and Biological Characteristics of Lake Erken

In this Case Study, we used the outputs from the transient climate model (tRCM) to drive the biogeochemical model (BIOLA) to simulate the long-term change in the physical, chemical and biological characteristics of Lake Erken, one of the most intensively studied lakes in Sweden. The model was forced with discharge and

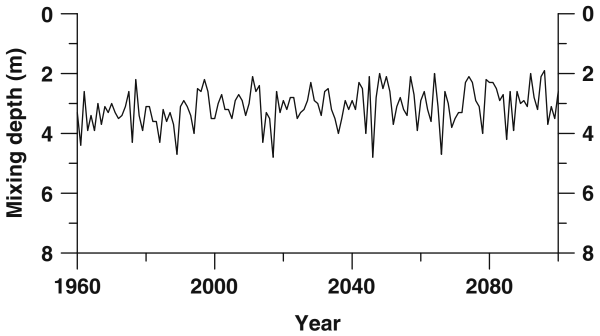

Fig. 15.4 The simulated inter-annual variation in the mean mixing depth for August for the period 1961–2100

loadings simulated by a simple runoff model linked to the tRCM. These outputs included a representation of both the seasonal and long-term variations in the hydrology but we assumed that there would be no systematic change in land-use.

The most striking result of the simulation was the progressive reduction in the ice-cover and the associated increase in the average temperature. A general decrease in the duration of ice-cover has recently been reported in a number of European lakes (see Chapter 4, this volume). In Lake Erken, the period of ice cover decreased from an average of two months in 1990–1999 (Chapter 7, this volume) to an average of only two weeks in 2071–2100. During the same period, the average annual water temperature increased by 1.2°C there was a small reduction in the mixing depth predicted in late summer (Fig. 15.4). These changes, coupled with the projected increase in primary production, led to, a substantial reduction in the summer hypolimnetic oxygen concentration (Fig. 15.5), a longer period of anoxia and a major increase in the area and volume of water affected. All these changes are likely to enhance the diffusion of phosphate from the sediment into the water and increase the amount entrained from the hypolimnion into the epilimnion.

15.3.3 Case Study 3: The Impact of Climate Change on the Growth of Phytoplankton in Esthwaite Water

The this Case Study, we used the PROTECH model to simulate the potential effects of climate change on the seasonal variations in the biomass of phytoplankton in Esthwaite Water, a small lake in the UK. Esthwaite Water is one of the most productive lakes in the English Lake District. It covers an area of 1.01 km^2 and has a mean depth of 6.4 m, a maximum depth of 15 m and a mean residence time of 0.25 years. The model was forced with the Hadley Centre HadAM3 regional climate model (pRCM) using the A2 and B2 scenarios suggested by the IPCC. The catchment model used to generate the seasonal flux in the supply of nutrients was

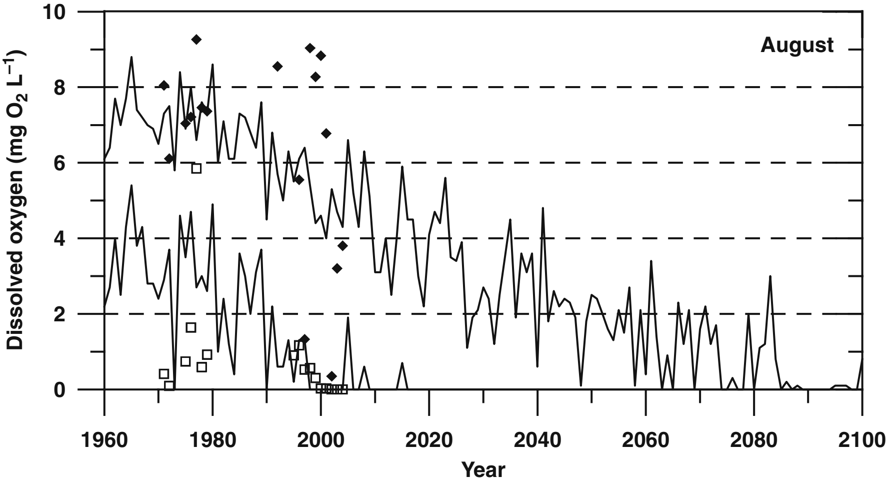

Fig. 15.5 The simulated (*lines*) and observed (*dots*) variation in the mean oxygen concentration in Lake Erken in August at 10.5 m (*diamonds*) and 19.5 m (*squares*) depth for the period 1961–2100

GWLF. Further details of the formulation and the calibration procedures for the GWLF model are given in Chapter 3.

The results (Fig. 15.6) show that the model predicts the essential features of the growth cycle under future as well as current climatic conditions. The most obvious difference between the control and future simulations are the predicted reductions in the early summer biomass of phytoplankton. The most likely explanation for this reduction is the projected reduction in the supply of phosphorus from the surrounding catchment.

15.3.4 Case Study 4: The Increased Incidence of Cyanobacterial Blooms in Galten

In this Case Study, we combined the outputs from pRCM. PROBE, GWLF and a variant of PROTECH (PROTBAS) to explore the potential effects of changes in the climate on the growth of cyanobacteria in the most productive basin of Lake Mälaren, Galten. In this sequence, the pRCM outputs were used to drive both the catchment model and the lake models to produce predictions for the supply of nutrients, ice cover, the temperature of the water and the mixing characteristics of the lake. These results were then used to predict the seasonal variations in the total biomass of phytoplankton and the biomass of buoyant cyanobacteria, i.e. *Anabaena*, *Aphanizomenon* and *Microcystis*. The models were calibrated with 21 years of observational data (1970–1990) and perturbed by the pRCM outputs for 2080–2100 using the IPCC ‘A2’ scenario.

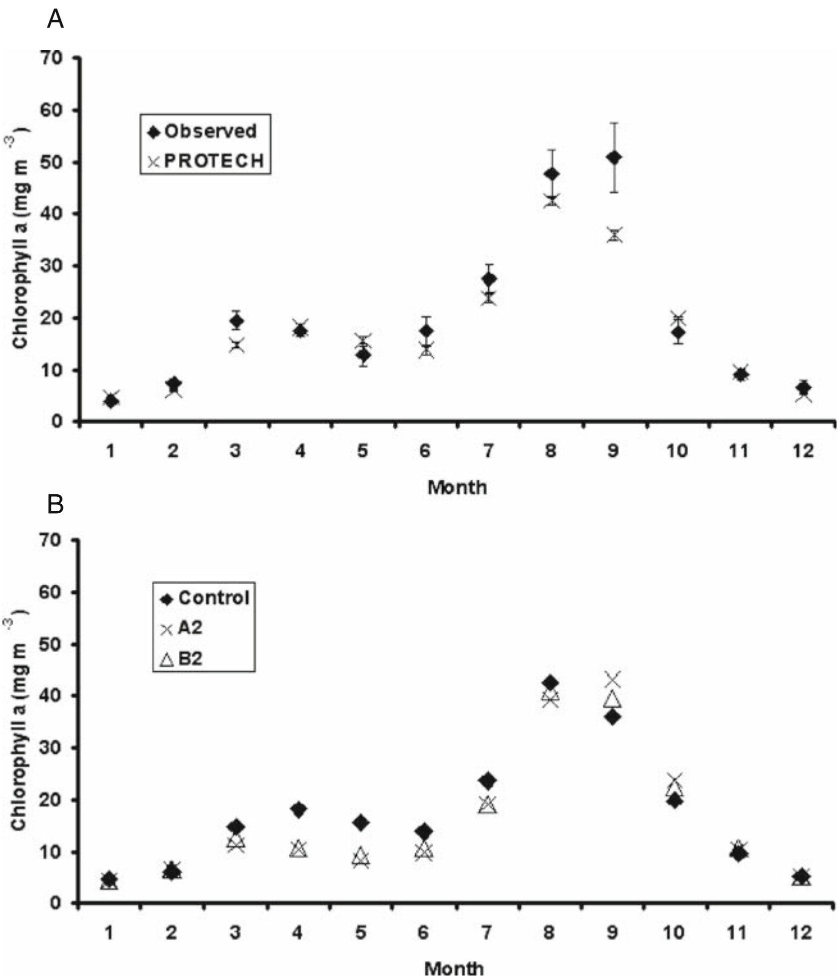

Fig. 15.6 The seasonal (monthly) variation in the mean biomass of phytoplankton in Esthwaite Water under current (**A**) and future (**B**) climatic conditions. The 'control' results are compared with the monthly means for data acquired between 1961 and 1990. The future simulations are based on the 2071–2100 outputs for pRCM using the IPCC 'A2' and 'B2' scenarios

The results generated for the future climate (Fig. 15.7) suggest a slight increase in the total phytoplankton biomass with an associated shift towards a dominance of buoyant, nitrogen-fixing cyanobacteria at the expense of diatoms. The increase in the annual mean cyanobacteria biomass was almost 50%, from 3 $\mu g\ l^{-1}$ (control period) to 4.5 $\mu g\ l^{-1}$ (scenario). The main factor responsible for these differences was the projected change in the hydrology of the catchment, i.e. higher streamflow during the whole winter period and lower streamflow during the growing season (May−October). The associated nutrient load was consequently lower during the

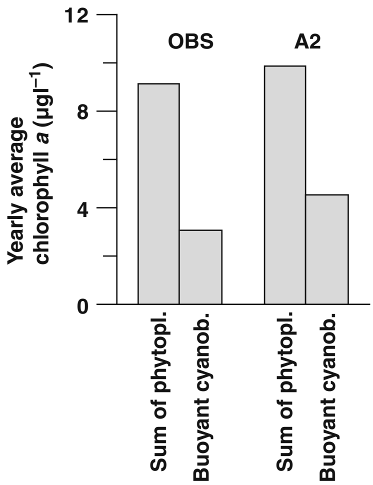

Fig. 15.7 The projected change in the mean chlorophyll concentration and the biomass of buoyant cyanobacteria in the Galten basin of Lake Mälaren. The values labelled 'OBS' are for the 1970–1990 'control' simulations and those labelled 'A' for the 2071–2100 scenario

growing season, which reduced the diatom biomass and increased the biomass of the nitrogen-fixing cyanobacteria. In this simulation, changes in the patterns of stratification in this polymictic lake had no affect on cyanobacteria biomass. This suggests that the most important climatic effects were those connected with the supply of nutrients and that the enhanced growth of cyanobacteria was primarily related to their nitrogen fixing capacity rather than their ability to regulate their depth in the water column.

15.4 Discussion

In this chapter, we assembled some examples to show how a variety of lake and catchment models can be linked to quantify the climatic responses of a number of different lakes. The results demonstrate that, by combining such models, we can generate outputs that quantify the indirect as well as the direct effects of the changing climate on the dynamics of the lakes. The results also show that many of these effects are very lake specific. The main 'water quality' effects projected from the four case studies are: (a) an increase in the total phosphorus loading (b) a decrease in the concentration of oxygen in deep water (c) a modest increase in the total biomass of phytoplankton and (d) a more pronounced increase in the abundance of cyanobacteria. Furthermore, the Case Studies showed that the climatic sensitivity of the individual lakes was critically dependent on their mixing characteristics, the climate-induced change to their nutrient loads and the residence times.

In these Case Studies, we have used a cascade of models of widely differing complexities to provide the 'water quality' projections required by the lake and catchment managers. At present, this 'linked mode' approach provides one of the most effective ways of assessing the potential impacts of climate change on water resources throughout Europe. At present, we cannot construct a single model to simulate all these complex processes so new ways have to be developed to integrating the outputs from existing formulations. In Europe, a promising new initiative has been the development of the Open Modelling Interface (Open MI) project (Blind et al., 2005). The Open MI software allows modelers to select and combine the most appropriate modeling tools available and test the sensitivity of their chosen approach. At present, Open-MI is best regarded as a demonstration project but supporting material is available under an Open Source license and is freely downloadable (www.openmi.org).

One important issue currently being addressed is the uncertainty associated with these 'linked' model results, i.e. the results have to encapsulate the effects of several different climate change scenarios. In most cases, there are at least four different sources of model uncertainty. Firstly, the uncertainty in the observational data required to drive the models. These data are typically the result of quite intensive sampling but there are still uncertainties associated with the measurement techniques and the spatial and temporal resolution of the surveys. For example, mass-balance nutrient models are often driven by single measurements on the main inflows, which, in ungauged inflows, are strongly influenced by antecedent conditions. Secondly, there is uncertainty in the structure of the models and their state parameters. These parameters are often very difficult to measure in the field so values are either assumed or taken from experiments run under controlled conditions. Interactions of the different parameters in the model are often complex and may the parameters may compensate each other due to the interaction of different processes. Thirdly, the extension of these models to future climatic conditions involve assumptions that are also uncertain (Pappenberger and Beven, 2006). Using the outputs from any climate model to drive any environmental model introduces further uncertainties that are very difficult to minimize. In climate change research, it is very important to recognise that the results produced by any model are a projection of a specific case and not a 'universal' prediction. Models based on different climate scenarios inevitably produce different results so ensembles of these simulations have to be assembled to provide a reasonable measure of likely outcomes. Fourthly, it is important to realize the uncertainty associated with the observations used to calibrate and validate the models. If the variability connected with some of these variables is large, many samples must be analyzed to reduce the confidence intervals for the calculated averages. It is now clear, that a number of practical and organizational problems need to be addressed before we can improve our assessments of the likely impact of climate change on water quality. One aim should be to construct and compare models of biogeochemical fluxes and ecosystem function for a range of different lakes and catchments situated in diverse landscapes. Here, the CLIME project has made a start by applying the same group of models to a number of lakes situated in different climatic regions (see applications by Persson et al., 2005 and Malmaeus et al., 2005). These studies have demonstrated that it is possible to apply the same generic models to a range of sites and use these results to quantify their climatic sensitivity. In future, more attention should be paid to the application of these models in more extreme environments and in situations where extreme climatic events are known to be important, e.g. the very high summer temperatures experienced over much of Europe 2003 (Schär et al., 2004). This is important because many of the parameters only valid over the range of conditions used to calibrate the models. When these conditions change in non-linear or step-wise way, the model outcome cannot be reliable. All the CLIME simulations were based on the simplifying assumption that there would be no systematic change in the regional land-use. In practice, we know that these land-use practices will change and be influenced by a variety of political and economic as well as climatic factors. Here, new and innovative approaches are needed to understand processes controlling the biogeochemical linkages between terrestrial and aquatic ecosystems. Although this is an old issue (see for example Likens et al., 1970) there is still a need to quantify and model these linkages. There have been significant advances in modelling of both terrestrial and aquatic biogeochemical processes, however, innovative approaches are needed to relate terrestrial, hydrological and biogeochemical processes to aquatic ecosystem function.

To further improve model projections, more spatially and temporally intensive monitoring is required. The nature of ecosystems is characterized by their high variability in space and time which is still not fully understood (Hughes et al., 2005; Humbert and Dorigo, 2005). The main problem in this respect is the availability of data (Porter et al., 2005). Many national organizations are reducing both the intensity and range of their environmental monitoring programs. However, with the growing realization that the uncertainty associated with these measurements more data are required to improve future predictions. One way to archive this is to install new automatic monitoring stations. The continued development of more sophisticated and more affordable sensors and communication systems means that such instruments can now be used to measure a wide-range of water-quality attributes in remote environments (Rouen et al., 2000; Porter et al., 2005). The use of these automatic stations will, however, dramatically increase the volume of ecological data generated and provide modellers with new opportunities to integrate their formulations over a range of different scales.

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Chapter 16
The Influence of Changes in the Atmospheric Circulation on the Surface Temperature of Lakes

Glen George, Ulrike Nickus, Martin T. Dokulil, and Thorsten Blenckner

16.1 Introduction

Seasonal and inter-annual variations in the circulation of the atmosphere have a major effect on the physical dynamics of lakes. In winter, these variations regulate the surface temperature of the lakes and their ice phenology (Livingstone, 2000; Livingstone and Dokulil, 2001; Chapter 4, this volume). In summer, they influence the mixing characteristics of the lakes and their physical stability (George and Taylor, 1995; George, 2006). Climatologists use a number of different methods to quantify these variations in the circulation. Some are based on large-scale features of the atmosphere, like the pressure gradient associated with the North Atlantic Oscillation (Hurrell, 1995). Others describe the variations observed at smaller scales using the weather typing techniques developed in synoptic climatology. Synoptic climatology has formally been defined as the study of the relationship between the atmospheric circulation and the local or regional climate (Barry and Perry, 1973). Yarnal (1993) and Yarnal et al., (2001) have produced reviews of developments in the field whilst Bardossy and Caspary (1990) have summarized some of the main circulatory changes recorded in Europe since the nineteenth century.

In CLIME, we used three different systems of regional weather typing to explore the effect of year-to-year variations in the weather on the thermal characteristics of lakes. In this chapter, we explain how these techniques were used to explain the surface temperature variations recorded in a number of lakes located in Northern, Western and Central Europe. Further information on the physical, chemical and biological characteristics of these lakes can be found elsewhere in this volume (e.g. Chapters 17, 18 and 19). Here, we describe the different systems used to classify the weather and show how the seasonal and inter-annual variations in the frequency of these weather types influence the observed variations in the surface temperature of the lakes. Weather type classifications of this kind have been widely used for

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hydrological impact assessments (Kilsby et al., 1998; Wilby et al., 1997) but the only limnological studies published to date are those of Blenckner and Chen (2003) in Sweden and George et al. (2007a, b) in Britain and Ireland.

16.2 Classifying the European Weather

The classification schemes devised for the European region cover a wide range of spatial scales. A summary of these methods was published by Lamb (1972) whilst Jacobeit et al. (2001) have described some of the changes recorded in the region since the eighteenth century. All these methods are based on the analysis of the spatial variations in the atmospheric pressure measured at a specified height. The early systems were based on the subjective analysis of these patterns on individual weather charts. Modern systems use semi-automated or automated techniques to arrive at the same synoptic categories. In CLIME, we used three, closely related, systems of classification to describe the synoptic situations that develop in Northern Europe, Western Europe and the Alpine region of Central Europe.

The Northern European scheme was designed by Chen (2000) and is based on pressure data acquired from an area that extends from 52.5° to 72.5°N and 5.0° to 27.5°E. Within this area, the synoptic situation is classified using measures that quantify the zonal and meridional pressure gradients, the geostrophic wind speed and the sheer vorticity. The scheme includes twenty seven circulation types but four types (anticyclonic, cyclonic, westerly and south westerly) account for more than 60% of the observed variation. The Western European scheme (Jenkinson and Collison, 1977) is an automated version of the subjective system developed by Lamb (1950). This scheme includes eleven circulation types but three types (westerly, anticyclonic and cyclonic) account for more than 50% of the recorded variation. Jones et al. (1993) have compared the results produced by the automated and subjective procedures and shown that there is a very high correlation between the two methods. The Central European system was developed by Steinacker (2000) to describe the weather patterns observed in the Alpine region of Austria. This system is based on ten circulation types, eight of which refer to the direction of flow, one is allocated to days when the pressure gradient is weak and one to days when the area is dominated by a major front. For practical reasons, we use slightly different methods to summarize the inter-annual variations observed in the three regions. In Northern Europe, we used counts of the ‘pure’ weather types described by Chen (2000). In Western Europe, we used indices based on weighted averages of the ‘pure’ and ‘hybrid’ frequencies, where the ‘pure’ weather types scored 2 and the ‘hybrid’ types scored 1 (Kelly et al., 1993). In Central Europe, we used counts of the ‘pure’ frequencies for most of the weather types but the northerly values were based on the sum of all categories with a northerly component. To simplify the inter-regional comparisons, we describe each index in the same way: where the first part of the name identifies the system of classification and the second part the type of flow e.g. Chen Anticyclonic, Lamb Westerly etc. Where possible, the examples selected, have been

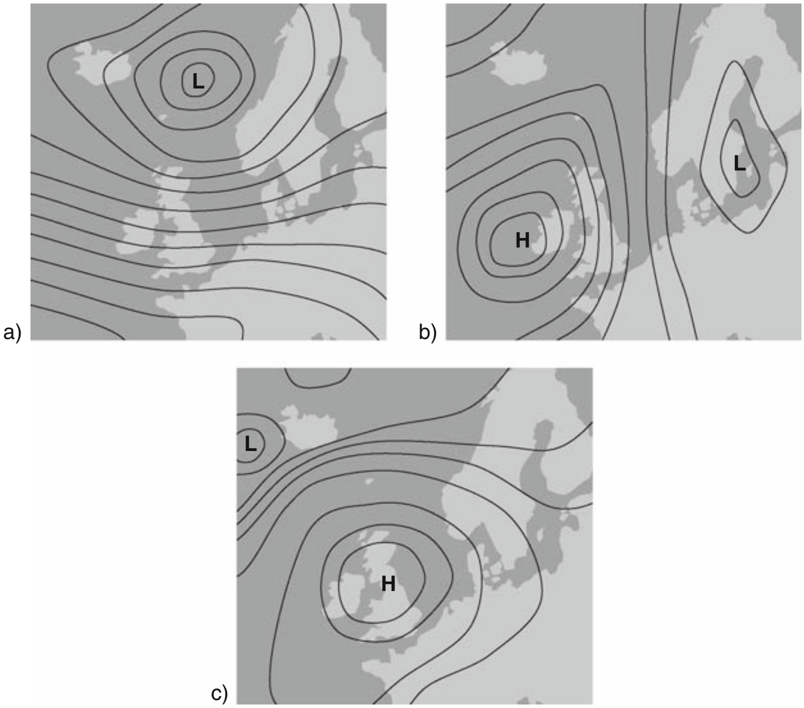


Fig. 16.1 The pressure distributions associated with the three basic patterns of flow, shown for an area centered on Britain and Ireland. (**a**) Zonal circulation. (**b**) Meridional circulation. (**c**) Rotational circulation. The letters show the position of the high (H) and low (L) pressure cells

designed to illustrate the effects associated with the three fundamental patterns of flow: the zonal, the meridional and the rotational.

The maps in Fig. 16.1 show the pressure distributions associated with these patterns for a geographic area centred on Britain and Ireland. Each weather type is classified according to the location of the high and low pressure cells and the mass flow of air over the region. In the zonal category (Fig. 16.1a), the flow is aligned with the west-east axis and there is usually a region of high pressure to the south and a region of low pressure to the north. This situation is commonly associated with changeable weather, mild winters, cool summers and higher than average wind speeds. In the meridional category (Fig. 16.1b) the flow is aligned with the north−south axis and there is usually high pressure in the west and low pressure to the north-east. This situation can give rise to cold weather at any time of year and is often associated with ice and snow during the winter. The rotational category (Fig. 16.1c) includes the common situation where a mass of air is moving clockwise around a stable area of high pressure. In Britain and Ireland, George et al. (2007b) have shown that these anticyclonic conditions have a major effect on the thermal characteristics of the lakes. In winter, this synoptic situation typically results in cold, clear conditions and an increased risk of frost. In summer, it is associated with warm, dry weather and below average wind speeds. Similar maps can be drawn to illustrate the synoptic situations that develop in the other European regions. In each case, the key attribute is the direction of the mass flow of air over the region of interest. For example, in northern Sweden, the situation shown in Fig. 16.1a would be classified as 'southerly' since the mass flow of air in that area is from south to north.

16.3 The Seasonal Variation in the Different Weather Types

In this section, we summarize the inter-annual variations in the weather types recorded in the three regions using long-term averages of their monthly frequency. Fig. 16.2 shows the seasonal variation in the weather types that dominated the three regions between 1960 and 2000. In Northern Europe, the dominant weather type was the anticyclonic (A) and the highest frequencies were recorded in the first five months of the year. The westerly weather type (W) was more common during the winter but there was no seasonal variation in the frequency of the northerly weather type (N). In Western Europe, the pattern of variation was very similar to that recorded in Northern Europe. The dominant weather type was the anticyclonic but the highest frequencies were now recorded in late spring and in early summer. The seasonal variation in the frequency of the westerly type was also more pronounced with very high frequencies recorded during the winter. Here, the frequency of the northerly weather type was lower and there was no appreciable variation in the seasonal totals. In Central Europe, the seasonal variation in the dominant weather types was different from that observed in the other regions. In the Steinacker system,

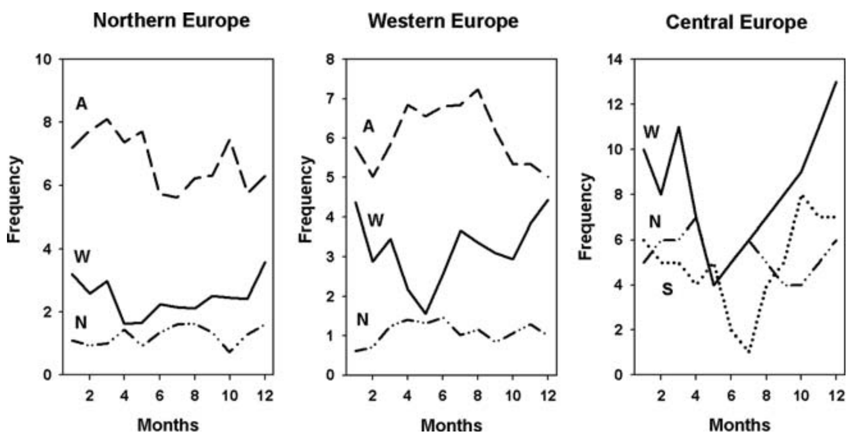


Fig. 16.2 The seasonal variation in the dominant weather types recorded in the three regions between 1960 and 2000. A = Anticyclonic, W = Westerly, N = Northerly and S = Southerly. (all frequencies in days per month)

the only attribute considered is the direction of movement and there is no record of the number of days with stationary ‘rotational’ conditions. In this region, the most important directional categories were the westerly and the northerly types but the southerly type was also quite common. The seasonal variation in the westerly type was very similar to that recorded elsewhere but the northerly frequencies were more variable and there was a pronounced summer decrease in the frequency of the southerly type. This circulation type is commonly associated with the north-south movement of the pressure fronts that are known to have a major effect on the weather experienced in the mountains.

16.4 The Lake Temperature Data Analyzed in the Three Regions

Table 16.1 summarizes the physical characteristics of the seven lakes that were used in the regional comparisons. The seven sites were physically very different but each lake was broadly representative of the type found in that particular region. The shallowest lake was Esthwaite Water, a eutrophic lake in the English Lake District. The deepest lake was Traunsee, an oligotrophic lake situated in the Saltzkammergut region of Austria. Most of the lakes were either free of ice during the winter or were only frozen for a few weeks in the year. The only exception was Lake Erken, a relatively shallow lake which is covered with ice for at least three months in the year. All the lakes were sufficiently deep to stratify during the summer but there were large site-to-site variations in the intensity of stratification and the depths of the seasonal thermoclines.

The temperature records for the seven sites are amongst the most extensive available in Europe. The surface temperatures of the British and Austrian lakes have been monitored for more than 40 years but the Swedish records are a little shorter. At all sites, measurements were taken at a central site using a temperature sensor with an accuracy of at least 0.1°C. Measurements were typically taken at weekly or fortnightly intervals with occasional gaps in severe winters. In the analyses reported here, we have combined all the available measurements to produce seasonal averages of the surface temperature that are then correlated with the frequencies of the dominant weather types. In each region, the winter months were considered to be December, January and February; the spring months March, April and May; the

Table 16.1 The physical characteristics of the lakes used in the weather type analyses

Lake	Region	Area (km^2)	Altitude (m)	Mean depth (m)
Erken	Northern Europe	11.1	11	9.0
Vänern	Northern Europe	5,655	27	27.0
Esthwaite Water	Western Europe	1.0	65	6.4
Windermere (North)	Western Europe	8.1	45	25.1
Mondsee	Central Europe	14.2	481	36.0
Hallstättersee	Central Europe	8.6	508	64.9
Traunsee	Central Europe	25.6	422	89.7

summer months June, July and August and the autumn months September, October and November. Both the raw time-series and the residuals in the fitted regressions were checked for serial correlation and for the presence of extreme outliers. Since only a small proportion (less than 5%) of the serial correlations were statistically significant these values were attributed to chance.

16.4.1 The Response of the Lakes to a Zonal Circulation

In all three regions, the dominant zonal flow was from west to east and the highest frequencies of this westerly type were invariably recorded during the winter. Figure 16.3 shows the relationship between the winter characteristics of three CLIME lakes and the relevant westerly index. In Northern Europe, the ice-break date on Lake Erken was negatively correlated with the Chen Westerly Index (r = -0.52) and the fitted regression was statistically significant at the 0.1% ($p < 0.001$) level. The difference between the earliest and the latest ice-break dates was 89 days and the earliest break-up date was that recorded in 1990. In Western Europe, the winter temperatures of the North Basin of Windermere were positively correlated with the Lamb Westerly Index ($r = 0.61$) and the fitted regression was statistically significant at the 0.1% ($p < 0.001$) level. The difference between the temperatures recorded in the winters with the lowest and highest indices was 2.4°C and the highest winter temperature was that recorded in 1998. In Central Europe, the surface temperatures of Traunsee were positively correlated with the Steinacker Westerly Index ($r = 0.51$) and the fitted regression was statistically significant at the 0.1% ($p < 0.001$) level. The difference between the temperatures recorded in the winters with the lowest and highest Westerly Index was 3.1°C and the highest winter temperature was that recorded in 1995.

In Europe, the winter frequency of westerly conditions is closely correlated with the atmospheric pressure gradient known as the North Atlantic Oscillation (Hurrell,

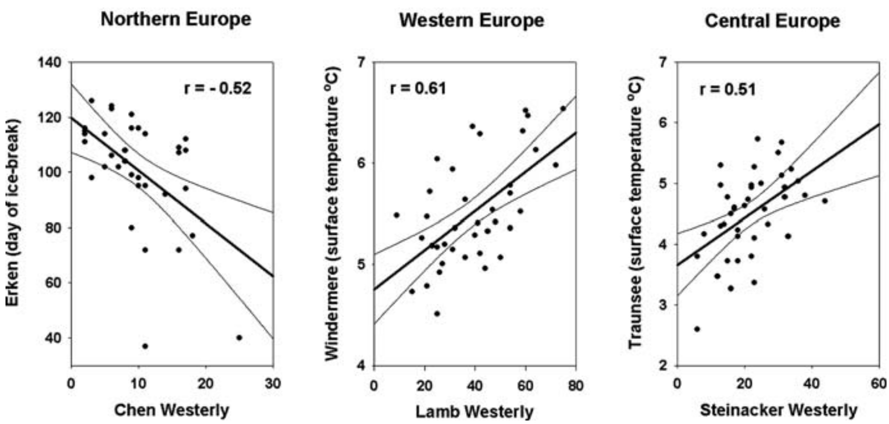


Fig. 16.3 The influence of westerly weather on the winter characteristics of the lakes

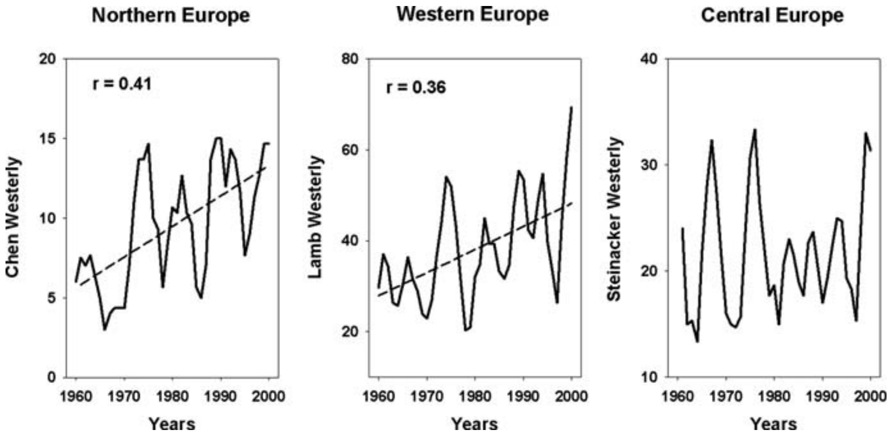


Fig. 16.4 The inter-annual variations in the winter Westerly Indices calculated for the three regions

1995). When the North Atlantic Oscillation (NAO) is in its positive phase, there is a strong westerly flow of air over the North Atlantic and winters in the northern half of Europe are milder and wetter. The pressure variations quantified by the Westerly Indices are regional manifestations of this north–south gradient. Figure 16.4 shows the year-to-year variations in the winter westerly indices calculated for the three regions between 1960 and 2000. The linear regressions fitted to the Chen and Lamb time-series were statistically significant at the 0.2% ($p < 0.002$) level but there was no long-term trend in the Steinacker time-series.

16.4.2 The Response of the Lakes to a Meridional Circulation

In all three regions, the dominant meridional circulations were those associated with the northerly weather type. The frequency of this weather type was often close to that of the southerly type but the northerly situation had a more pronounced effect on the temperature of the lakes. Figure 16.5 shows the relationship between the autumn surface temperature of three CLIME lakes and the frequency of these northerly conditions. In Northern Europe, the surface temperature of Vänern was negatively correlated with the Chen Northerly Index ($r = -0.38$) and the fitted regression was statistically significant at the 2% ($p < 0.02$) level. The difference between the temperatures recorded in the autumns with the lowest and highest indices was 3.6°C and the highest autumn temperature was that recorded in 1997. In Western Europe, the surface temperatures of Esthwaite Water were also negatively correlated with the Lamb Northerly Index ($r = -0.66$) and the fitted regression was statistically significant at the 0.1% ($p < 0.001$) level. The difference between the temperatures recorded in the autumns with the lowest and highest indices was 3.6°C and the highest autumn temperature was that recorded in 1995. A very similar pattern was recorded at the site selected from Central Europe. Here, the autumn temperature of Hallstättersee

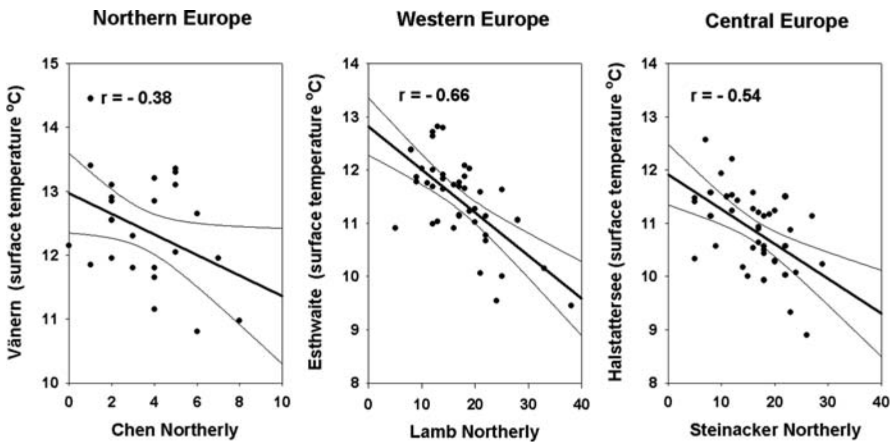


Fig. 16.5 The influence of northerly weather on the autumn temperature of the lakes

was negatively correlated with the Steinacker Northerly Index ($r = -0.54$) and the fitted regression was statistically significant at the 0.1% ($p < 0.001$) level. The difference between the temperatures recorded in the autumns with the lowest and highest indices was 3.7°C and the highest autumn temperature was that recorded in 1961. At each location, the strength of the correlation was closely related to the observed difference in the air and water temperatures. Thus the strongest negative correlation was recorded in Western Europe where the movement of cold air from the north increased the rate at which heat was lost from the lake during the autumn.

Figure 16.6 shows the year-to-year variations in the autumn Northerly Indices calculated for the three regions between 1960 and 2000. In Northern and Western

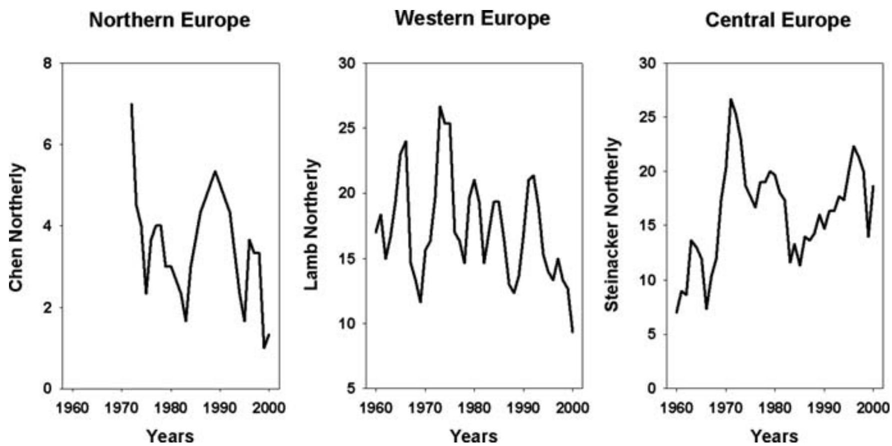


Fig. 16.6 The inter-annual variations in the autumn Northerly Indices calculated for the three regions

Europe, this frequency decreased over the period of observation but the pattern was quite the reverse in Central Europe. Fitting linear regressions to these variations showed that the trends were not statistically significant. The trend observed in Central Europe appears to be more sustained than that recorded in the other regions, a feature that is thought to reflect a systematic shift in the position of the frontal systems that are very common at this time of the year.

16.4.3 The Response of the Lakes to a Rotational Circulation

The Steinacker system was designed to quantify the directional movement of air and does not include any rotational category. In Northern and Western Europe, the rotational category that had the most pronounced effect on the temperature of the lakes was the anticyclonic type. Figure 16.7 shows the relationship between the summer surface temperature of two CLIME lakes and the changing frequency of the rotational weather type. In Northern Europe, the surface temperature of Vänern was positively correlated with the Chen Anticyclonic Index ($r = 0.62$) and the fitted regression was statistically significant at the 0.1% ($p < 0.001$) level). The difference between the temperatures recorded in the summers with the lowest and highest anticyclonic index was 6.6°C and the highest temperature was that recorded in 1997. In Western Europe, the summer temperature of the North Basin of Windermere was positively correlated with the Lamb Anticyclonic Index ($r = 0.77$) and the fitted regression was statistically significant at the 0.1% ($p < 0.001$) level). The difference between the temperatures recorded in the summers with the lowest and highest index was 3.6°C and the highest temperature was that recorded in 1997. Figure 16.8 shows the year-to-year variations in the summer rotational indices calculated for the two regions between 1960 and 2000. In both cases, there were large year-to-year variations but there was no significant long-term trend.

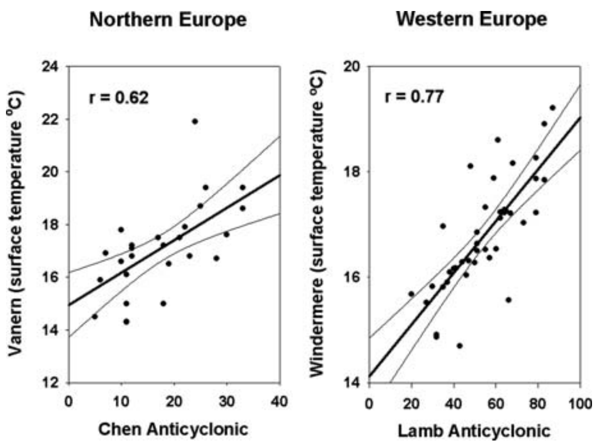


Fig. 16.7 The influence of anticyclonic weather on the summer temperatures of the lakes

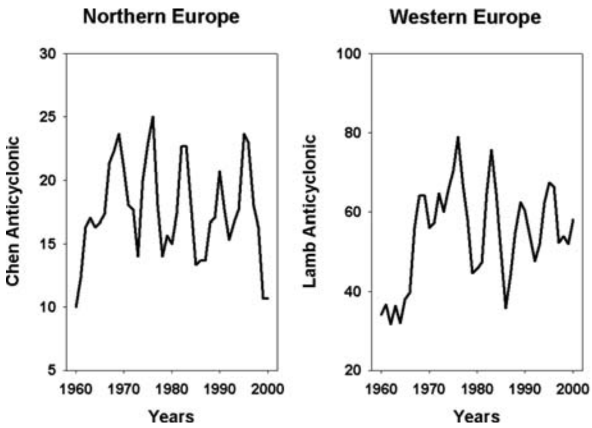


Fig. 16.8 The inter-annual variations in the summer Anticyclonic Indices calculated for Northern and Western Europe

16.5 Relationships Between the Regional Weather Types and the General Circulation

The conditions quantified by the weather type indices are regional manifestations of much larger scale features in the circulation of the atmosphere (Spellman, 1997; Chen and Hellström, 1999). The strongest links are with the North Atlantic Oscillation (NAO), a feature that dominates the movement of air over a large part of the Northern Hemisphere (Hurrell et al., 2003). Figure 16.9a shows the relationship between the Lamb Westerly Index for the winter period and the NAO Index used by Hurrell (1995). This index is based on the standardized difference in the atmospheric pressure recorded at Stykkisholmur in Iceland and Lisbon in Portugal. Positive values of the index are associated with stronger than average westerly flows, a feature reflected in the high frequency of westerlies recorded over Britain and Ireland. The calculated correlation between the Lamb Westerly Index and the NAOI was 0.79 and the fitted linear regression was statistically significant at the 0.1% ($p < 0.001$) level. Comparisons with the two indices have, however, shown that the weather type indices are better at explaining the temperature variations observed in the lakes. Figure 16.9b compares the proportion of the variation in the winter surface temperature of Windermere explained by the Lamb Westerly Index and the NAOI. In this analysis, correlation coefficients were calculated for successive ten-year 'blocks' of the surface temperature record. Thus, the correlation shown for 1960 is that calculated for the period between 1960 and 1969 whilst that shown for 1961 covers the period between 1961 and 1970. The results demonstrate that the two indices fluctuate in the same way but the Lamb index explains more of the observed inter-annual variation. The proportion of the variance explained by the NAOI was 28% but this increased to 38% when the independent variable was the Lamb index. In a similar way, a study by Blenckner and Chen (2003) in Sweden showed that the regional index developed by Chen (2000)

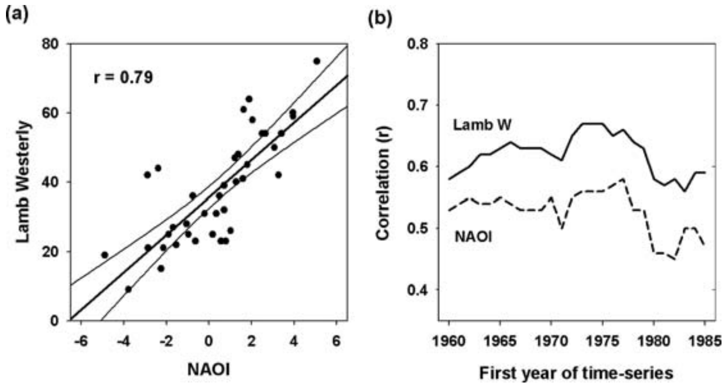


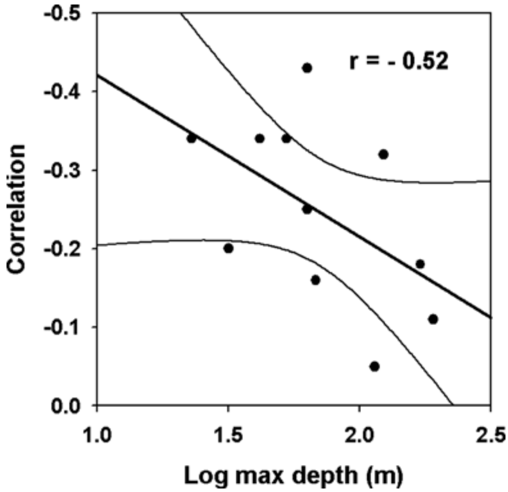
Fig. 16.9 (**a**) The relationship between the winter Lamb Westerly Index and the North Atlantic Oscillation Index (NAOI). (**b**) The extent to which the winter surface temperature of Windermere could be predicted using the Lamb Westerly Index and the NAOI

explained more of the observed variation in both ice-off and the timing of the spring bloom in Lake Erken than an index based on the NAO.

16.6 The Sensitivity of Different Types of Lakes to the Observed Variations in the Regional Weather

The response of any lake to a change in the weather is critically dependent on its size, its heat storage capacity and its exposure to wind mixing. Large lakes integrate the variations in the weather on longer time-scales whilst small lakes are more sensitive to short-lived events. In this chapter, we have selected examples to illustrate the climatic response of a range of different lakes. In Northern and Western Europe, the number of sites with complete temperature records was too low for any meaningful site-to-site comparisons. However, more data was available for the lakes in the Saltzkammergut area of Austria where long-term measurements are available for a number of lakes (Livingstone and Dokulil, 2001). Figure 16.10 compares the response of eleven of these lakes to the documented variation in the winter value of the Steinacker Northerly Index. These lakes are sufficiently deep to remain free of ice in most winters so the amount of heat lost during the winter is closely correlated with their maximum depth. In this analysis, we have used the correlation between the surface temperature of the lakes and the Steinacker index as a measure of their climatic sensitivity and then compared these correlations to their maximum depth. The maximum depths have been plotted on a logarithmic scale to normalize their distribution and a linear regression fitted to the observed trend. The trend in the correlations shows that the shallower lakes are more sensitive to the observed variations in the number of northerly days since they lose proportionately more of the heat stored during the summer. The strongest negative correlation was that recorded for Fuschelsee, a lake with a maximum depth of 65 m, and the weakest that recorded for Wolfgangsee, a lake

Fig. 16.10 The sensitivity of eleven Austrian lakes to the recorded variation in the winter weather. The correlations are those recorded between the surface temperature of the lakes and the winter value of the Steinacker Northerly Index (Data from Livingstone and Dokulil, 2001)



with a maximum depth of 114 m. The fitted linear regression explained 27% of the recorded variation and was statistically significant at the 5% ($p < 0.05$) level.

16.7 Discussion

Lakes are widely regarded as ideal integrators of the regional climate since they respond to the combined effects of the air temperature, the solar radiation and the wind (George, 2006). A number of investigators have explored the impact of large-scale changes in the atmosphere on the dynamics of lakes. Notable examples include the effects associated with the North Atlantic Oscillation (Straile et al., 2003; George et al., 2004) and the influence of the Gulf Stream on the atmosphere (George and Taylor, 1995; George, 2002). The weather type approach used here, explores these interactions on a scale that is much closer to that of the individual lakes. In the early days of synoptic climatology, most investigators simply described the schemes used and provided very few examples of their application. Today, these techniques are used for a variety of purposes, such as statistical downscaling (Conway and Jones, 1997), rainfall prediction (Linderson, 2001), plant phenology (Aasa et al., 2004), sea level prediction (Chen and Omstedt, 2005), local wind studies (Achberger et al., 2006) and the construction of regional climate change scenarios (Chen et al., 2006). In the freshwater field, the most common applications are those connected with the geographic variation in the rainfall (Wilby et al., 1997; Littman, 2000). The only limnological applications published to date are those of Blenckner and Chen (2003), George et al. (2007a) and George et al., (2007b). From a limnological point of view, these techniques have a number of practical benefits:

1. They are based on an analysis of a variable (atmospheric pressure) that is readily available and is less affected by local inhomogeneities than other meteorological variables (Jones et al., 1999).

2. They are synergistic measures that integrate the effects of a number of variables known to influence the dynamics of lakes.
3. They provide a useful historical perspective on the climatic changes observed in a particular region and can be used to compare these variations with those projected for a warmer world.
4. The regional indices are stationary but the areas influenced by large-scale features, such as the NAO, tend to change with the passage of time (see Omstedt and Chen, 2001).

16.7.1 Classifying the Weather in the European Area

Atmospheric pressure is one of the easiest meteorological variables to measure and long-term records are available for a number of sites distributed throughout Europe. The classification systems used here are all based on the same principles but the areas covered by the typologies vary from region to region. The Chen system covers a relatively large area of Northern Europe and was specifically designed to characterize the circulation patterns over Sweden. The Lamb system was designed to quantify the highly dynamic situations that develop over Britain but can also be used to highlight the mass movement of air over Ireland. The Steinacker system covers a much smaller area and was primarily designed to categorize the weather in the Alpine region. An important feature of these schemes is their ability to describe the synoptic situations that develop on a daily basis. They are thus particularly useful for investigating the responses of the lakes to short-term changes in the weather and assessing their sensitivity to extreme events. In practical terms, the limiting factor is the time required to characterize the changing meteorological situation i.e. the number of observations required to arrive at a statistically meaningful measure of the observed variation. In the examples presented here, the averaging period was three months but the same approach is currently being used to analyze the hourly measurements acquired by a network of automatic monitoring stations (Rouen et al., 2005). In a global context, it is worth noting that the methods used to classify the weather in Europe are different from those used in North America. In the US, such schemes are more commonly used to identify different air masses (Kalkstein and Nichols, 1996) in studies connected with effects of pollution (Comrie and Yarnal, 1992) or the impact of urban heat stress (Sheridan, 2002).

16.7.2 The Limnological Significance of the Observed Variations

A detailed analysis of the limnological effects associated with the climatic situations described here is outside the scope of this summarizing chapter. Some examples have been included elsewhere in this book (see Chapter 14 and Chapter 19). In this chapter, we have confined our analyses to the most immediate effects i.e. the impact on the surface temperature of the selected lakes. These variations are, however, closely correlated with more fundamental physical processes, such as the stability of the water column and the mixing characteristics of the lakes. Figure 16.11a

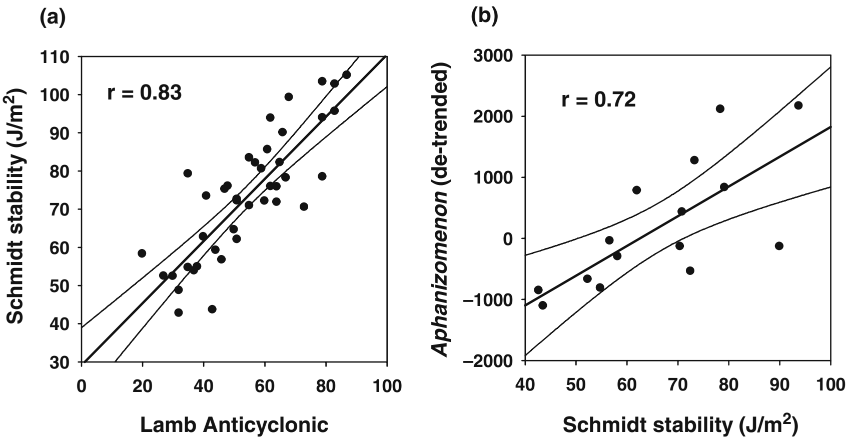


Fig. 16.11 (**a**) The relationship between the summer stability of Esthwaite Water and the Lamb Anticyclonic Index. (**b**) The influence of the stability of the water column on the de-trended abundance of Aphanizomenon in Esthwaite Water

shows the relationship between the average summer stability of Esthwaite Water and the Lamb anticyclonic index. The measure of stability used is that described by Schmidt (1928) and the fitted linear regression explains almost 70% of the observed inter-annual variation. In recent years, summers in the English Lake District have become both warmer and calmer (George et al., 2007a) as the number of anticyclonic days has tended to increase. This trend towards more stable conditions has already had a significant effect on the growth of phytoplankton with surface blooms of cyanobacteria becoming more common (Paerl and Huisman, 2008). At one time, such blooms were thought to be an inevitable consequence of eutrophication but are now known to be strongly influenced by antecedent weather conditions (George et al., 1990; Huisman et al., 2004).

Figure 16.11b shows the effect that changes in the stability of the water column had on the growth of the blue-green alga *Aphanizomenon* in Esthwaite Water between 1960 and 1974. The *Aphanizomenon* numbers have been de-trended to compensate for the observed increase in the supply of nutrients and the residual variation related to the stability of the water column. The results show that *Aphanizomenon* numbers tend to increase whenever the summer is calm and there is a sustained increase in the stability of the water column. During the period shown here, *Aphanizomenon* was the dominant bloom-forming species and the fitted regression was statistically significant at the 1% ($p < 0.01$) level.

16.7.3 The Climatic Significance of the Observed Variations

Until recently, the only climate change scenarios available were those based on 'time slice' experiments. In these experiments, a regional climate model (RCM) was first run under historical conditions and then under the conditions projected for a warmer world. The historical simulations were based on the atmospheric conditions that existed between 1961 and 1990, the reference period recommended by the IPCC in 2001. The warm world simulations were based on the atmospheric conditions projected for 2071−2100 using boundary conditions defined by the latest General Circulation Models (GCMs). In 2005, the first results from a new series of 'transient' experiments were produced by the Swedish Meteorological and Hydrological Institute (Kjellstrom et al., 2005). In these experiments, the RCMs were run for a continuous period that extended from 1960 to 2100 to investigate how the climate might change from its present to a future state.

Figure 16.12 show the results produced by one of these simulations for the grid-box that covered the English Lake District. The results shown are for the Rossby Centre RCA3 model which is a development of the RCM's described by Samuelsson in Chapter 2. In this figure, the grey lines show the results from a transient experiment based on the IPCC 'B2' emission scenario and the black lines from an experiment based on the more extreme 'A2' scenario. The vertical bars show the range of winter and summer air temperatures recorded in the Lake District for the period between 1960 and 2000. Since the temperatures generated by the model were different from those observed, the bars have been centred on the average temperatures recorded in the area between 1960 and 2000. The results imply that, the temperatures projected by the model are unlikely to exceed the extremes recorded in the area for some considerable time. Such a conclusion has, however, to be treated with some caution since heat waves are becoming more common and the temperatures reported under these conditions are well outside this historical 'envelope'.

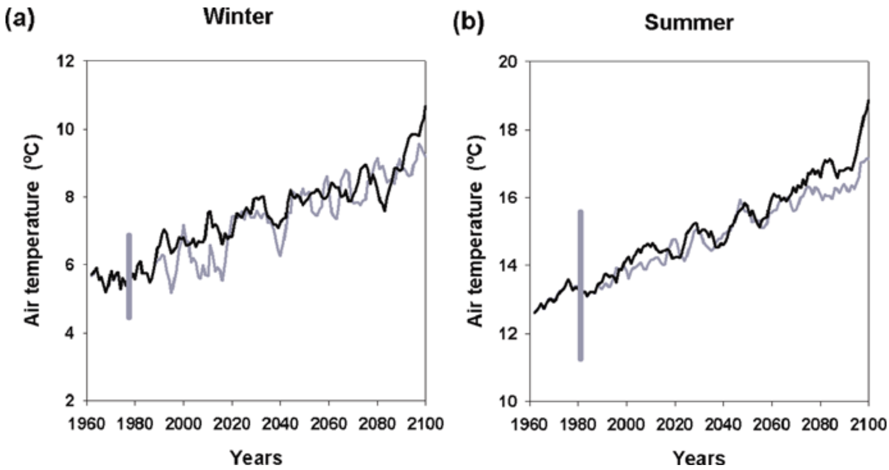


Fig. 16.12 The air temperature projections generated by the Rossby Centre transient model for the English Lake District in (**a**) winter and (**b**) summer. The *grey lines* show the results for the 'B2' scenario and the *black lines* the results for the'A2' scenario. The *vertical bars* show the range of the winter and summer temperatures recorded in the Lake District between 1960 and 2000

The weather typing approach described here provides a very effective means of identifying such extreme years and assessing their potential effect on the dynamics of lakes. For example, a reduction in the number of anticyclonic days recorded during the summer would significantly reduce the most immediate impacts of global warming. In contrast, a marked increase in the number of anticyclonic days would accentuate these impacts and amplify the changes observed in the most sensitive lakes. It is now widely recognised that the earth's climate system tends to operate in a small number of relatively stable configurations (Corti et al., 1999). In future, the weather patterns experienced in some parts of Europe could well change in a sudden way. The weather typing approach described here should help to quantify these tendencies and allow us to produce more robust assessments of their environmental impact.

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Chapter 17
Regional and Supra-Regional Coherence in Limnological Variables

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17.1 Introduction

Limnologists and water resources managers have traditionally perceived lakes as discrete geographical entities. This has resulted in a tendency for scientific lake studies to concentrate on lakes as individuals, with little connection either to each other or to large-scale driving forces. Since the 1990s, however, a shift in the prevailing paradigm has occurred, with lakes increasingly being seen as responding to regional, rather than local, driving forces (Livingstone, 2008). The seminal work on regional coherence in lake behaviour was that of Magnuson et al. (1990), who showed that many features of lakes within the same region respond coherently to drivers such as climate forcing and catchment processes. From this study it emerged that the degree of coherence among lakes is greatest for those properties most directly affected by climate forcing. Specifically, the physical properties of lakes tend to vary in a more coherent way than their chemical and biological properties (see also Kratz et al., 1998). Further overviews of the topics of coherence and climate-driven variability, focusing mainly on North American lakes, have been given by Magnuson et al. (2006a,b). In this chapter, we will examine the phenomenon of spatial coherence among time-series of some important physical, chemical and biological lake variables at regional and supra-regional scales in Europe. Here, spatial coherence is defined as the degree of correlation between time-series of measurements made simultaneously at different locations (in contrast to temporal coherence, which is defined as the degree of correlation between time-series of measurements made at one location but at different times). The concept of coherence in this context is not well-defined in a mathematical sense. It can be parameterised in various ways: for instance as the mean coefficient of determination (r^2) calculated between all possible lake pairs, or as the mean coefficient of determination calculated between each lake and the mean time-series of all other lakes (e.g., Livingstone and Dokulil, 2001), or

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as an intraclass correlation coefficient (e.g., Rusak et al., 1999). Regardless of the parameterisation employed, appropriate methods to remove the effect of daily and annual cycles were used in all examples mentioned in this chapter before computing coherence, so that, for instance, shared seasonal patterns among lakes do not affect the results. In the following, we use the term 'regional' to refer to spatial scales corresponding to a circle of about ten kilometres up to several hundred kilometres in diameter, and the term 'supra-regional' to refer to spatial scales any larger than this.

17.2 How Is a Climate Signal Transferred to Lakes?

Of the various ways in which a climate signal can be transferred to a lake, arguably the most important is via its effect on water temperature. The heat balance of a lake (Fig. 17.1) is primarily determined by five heat-exchange processes – three radiative and two non-radiative – that act at, or close to, the lake surface (Edinger et al., 1968; Sweers, 1976). These processes are: the absorption of short-wave solar radiation by the lake; the absorption of long-wave atmospheric radiation by the lake; the emission of long-wave radiation from the lake surface; the exchange of latent heat between the lake surface and the atmosphere due to evaporation and condensation; and the convective exchange of sensible heat between lake surface and atmosphere. These processes involve one essentially astronomically determined variable – clear-sky solar radiation – and four meteorological variables – air temperature, cloud cover, wind speed and relative humidity.

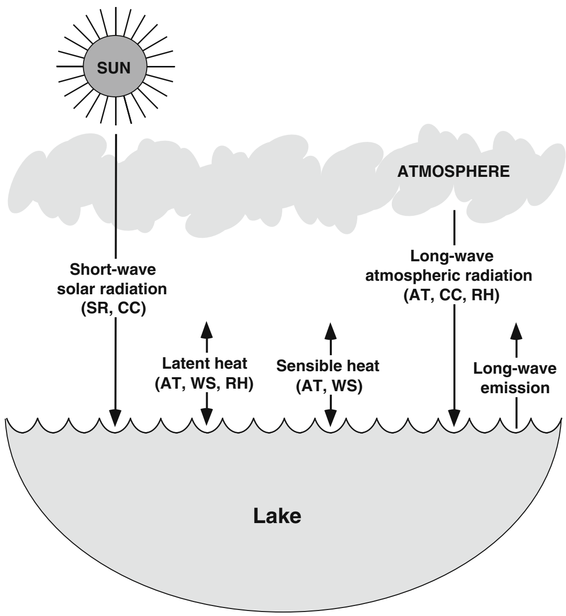

Fig. 17.1 Schematic diagram illustrating the five main heat-exchange processes that determine the heat balance of a lake. The main external variables involved are the essentially astronomically determined variable clear-sky solar radiation (SR), and the meteorological variables cloud cover (CC), air temperature (AT), wind speed (WS) and relative humidity (RH)

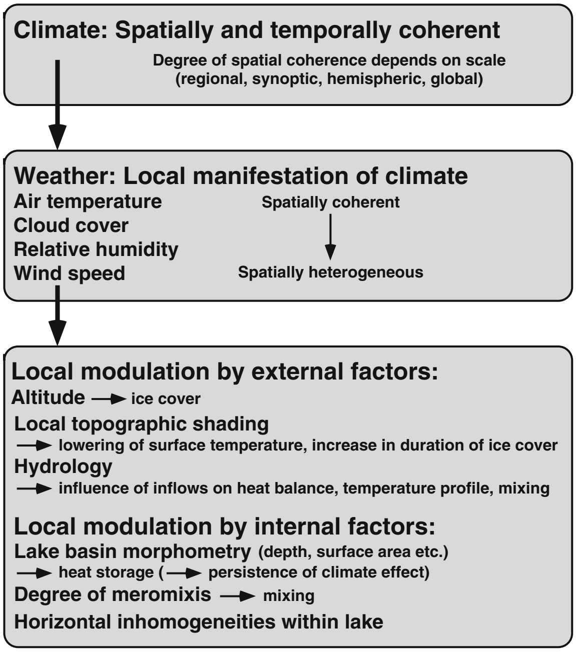

Fig. 17.2 An overview of the factors affecting the relationship between climatic forcing and the physical lake response to this forcing

At any given time, the surface water temperature of a lake tends exponentially towards an equilibrium temperature that is defined as the surface water temperature at which the net heat flux into the lake would theoretically be zero (Edinger et al., 1968). The equilibrium temperature is therefore driven by the same five external variables as the lake heat balance. The existence of spatial coherence in any of the five driving variables among the lakes of a given region would thus imply the existence of some degree of spatial coherence in the equilibrium temperatures of the lakes' surfaces, and hence in the lake surface temperatures themselves.

The link between climate forcing and the physical response of a lake to this forcing is illustrated in Fig. 17.2. Although the surface equilibrium temperature of any given lake is ultimately driven by local weather and not large-scale climate, climate and weather are tightly linked conceptually. On the one hand, climate can be viewed as the expression of a high level of spatial and temporal coherence in local weather. On the other hand, weather can be viewed as a local manifestation of climate, implying that local weather is comprised of a large-scale climate signal with superimposed local noise. The signal-to-noise ratio of the climate signal at a lake surface therefore depends on the relative importance of large-scale (e.g., synoptic) and small-scale (local) fluctuations. The individual meteorological variables that comprise the local weather exhibit varying degrees of spatial coherence; i.e., they have differing signal-to-noise ratios with respect to the climate signal. Clear-sky solar radiation, the ultimate astronomical driving variable, is responsible for the most obvious spatially coherent fluctuations in lake surface equilibrium temperature, and hence in lake surface temperature; viz. the diel and annual cycles. Of the four major meteorological variables that drive the lake surface equilibrium temperature, air temperature is highly spatially coherent (Jones et al., 1997), whereas cloud cover, relative humidity and wind speed are more spatially heterogeneous.

In addition to spatial heterogeneity in local weather, a further source of noise that can mask the climate signal at the lake surface is the modulation of local weather by factors related to lake location. Such factors include altitude and topographic shading, both of which affect lake surface water temperature and the duration of ice cover. They also include the hydrology of the inflows, which can affect the lake heat balance and the form of the temperature profile, and hence the frequency of occurrence and intensity of mixing. In mountainous areas, altitude above sea level can be a very important factor affecting the way lakes respond to climate forcing and the coherence of their response. With increasing altitude, lake surface water temperatures tend to decrease, mainly in response to the approximately linear decrease in surface air temperature (Livingstone et al., 1999, 2005a). In itself this need not necessarily affect the coherence of the physical response of lakes to large-scale climatic forcing; however, at high altitudes, surface water temperature can become increasingly decoupled from ambient surface air temperature, either as a result of a change in lake circulation regime or because of an increased influence of meltwater on the lake heat balance (Livingstone et al., 2005a). Altitude above sea level also determines the duration of ice cover (Eckel, 1955; Livingstone and Dokulil, 2001) and its spatial extent. Additionally, local topographic shading can modify the effects of altitude on lake surface water temperature by depressing it below the value expected due to altitude alone (Livingstone et al., 1999, 2005b; Goudsmit et al., 2000). Snow cover at high altitudes can affect the lake heat balance by feeding the inflow with a constant supply of water close to 0°C, thus decoupling it to some extent from climatic forcing (Livingstone et al., 2005a). Additionally, local variability in the presence or absence of snow on lake ice can affect water temperatures below the ice by influencing the degree of light penetration (e.g., Schindler et al., 1990). Altitude, topographic shading and local hydrology can thus play important roles in determining the strength of the primary physical response of lakes to climate forcing, and hence the magnitude of climate-related coherence among lakes.

In addition to modulation by external factors, internal factors can also affect the strength of the link between climate forcing and the physical lake response. Lake morphometry influences the distribution of heat, and can also influence the timing of ice-on (Stewart and Haugen, 1990) and the extent of vernal circulation (Salonen et al., 1984). Deep lakes tend to exhibit a more persistent physical response to climatic forcing than shallow lakes (Gerten and Adrian, 2001). The deep waters of lakes which are chemically stratified also tend to respond less sensitively to climatic forcing than the deep waters of lakes which are not (Livingstone, 1993, 1997). Large lakes – especially those with a convoluted coastline and multiple basins, such as the Swedish lakes Mälaren, Vättern and Vänern – can be exposed simultaneously to several different local weather regimes, to which they exhibit an internally heterogeneous response (e.g., Weyhenmeyer, 2004).

An important question arising from the above is: Are local effects strong enough to destroy the coherence imparted by the climate? If the answer to this question is yes, then surface water bodies are idiosyncratic individuals that merely produce noise, unrelated to any large-scale climate-related forcing. If the answer to this question is no, however, then surface water bodies can be viewed as local samples of a climatically-driven continuum. In the following, we will present evidence to support the latter hypothesis.

17.3 Short-Term Regional Coherence

Although lakes respond to local weather on many levels, their most direct response is physical. It is this direct primary physical response of lakes to meteorological forcing that is predominantly responsible for much of the indirect response, since weather-driven variations in the temperature profile, mixing regime and hydrology elicit secondary and tertiary responses at the chemical and biological levels. Thus, any regional coherence among lakes that might be detected on these secondary and tertiary levels is very likely to be the indirect result of coherence in the primary physical response. Because of the time-lags inherent in lake-internal processes, short-term coherence among lakes (i.e., on time scales of several days or less) is most likely to be detectable in the physical response. This response is strongest in the uppermost levels of the water column, and is most easily detectable in the surface water temperature.

Figure 17.3 illustrates some examples of short-term coherence in lake surface water temperature. Measurements made by Livingstone and Kernan (2009) in 25

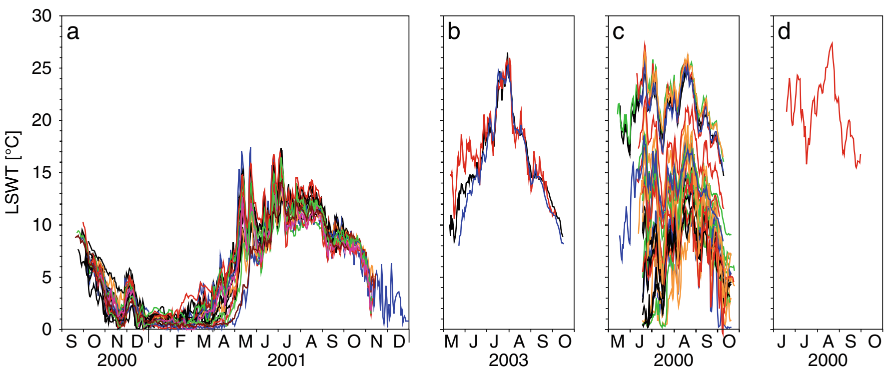

Fig. 17.3 Examples of short-term regional coherence in lake surface water temperature (LSWT). (**a**) LSWT measured in 25 lochs in the Grampians and the Northwest Highlands of Scotland in 2000 and 2001 (after Livingstone and Kernan, 2009). (**b**) LSWT measured in Võrtsjärv (Estonia), Peipsi (Estonia) and Pääjärvi (Finland) in 2003 (P. Nõges, T. Nõges, L. Arvola and M. Järvinen, unpublished data). (**c**) LSWT measured in 29 Swiss lakes in 2000 (after Livingstone et al., 2005a). (**d**) LSWT measured in Lake Balaton (Hungary) in 2000 (after Livingstone and Padisák, 2007)

lochs distributed throughout the Grampians and the Northwest Highlands of Scotland (Fig. 17.3a) show regional coherence in surface water temperature to be high during late spring and summer (when the epilimnion is thin), and autumn (when surface water temperatures are determined primarily by surface cooling). Coherence is substantially lower in winter, when fluctuations in lake surface temperature are small and the lochs may be partially ice covered, and in early spring, when they warm up and stratify at different times depending on their altitude and distance from the Atlantic. Measurements in Finland and Estonia show lake surface water temperatures to fluctuate extremely coherently during most of summer and early autumn (Fig. 17.3b). The lake surface temperatures illustrated in Fig. 17.3a,b not only fluctuate coherently within each region, they also differ very little in an absolute sense, making it relatively easy to obtain a reliable estimate of the surface temperatures of other lakes in the region concerned. However, this is only possible because there are no very large differences in altitude among the lakes: the Scottish lochs of Fig. 17.3a all lie within the range 720 ± 200 m a.s.l., and the Finnish and Estonian lakes of Fig. 17.3b lie within the range 67 ± 37 m a.s.l. In mountainous regions the situation is more complex. In the Swiss Alps below about 2,000 m a.s.l., lake surface water temperatures fluctuate coherently in response to short-term regional climatic forcing, reflecting fluctuations in regional air temperature (Livingstone and Lotter, 1998), but mean lake surface water temperatures decrease approximately linearly with increasing altitude (Livingstone et al., 1999). Figure 17.3c illustrates lake surface water temperatures measured in 29 lakes in Switzerland between 465 m a.s.l. and 2,470 m a.s.l. (Livingstone et al., 2005a). The large range of mean values is a result of the pronounced decrease in mean lake surface water temperature that occurs with increasing altitude; however, the degree of short-term coherence in lake surface water temperature among the lakes is still high despite the large differences in altitude. Livingstone et al. (2005b) showed that the pattern of deviations from linearity shown by the surface temperatures of lakes in the Swiss Alps varies little from month to month and from year to year, implying that most local effects do not substantially diminish the short-term coherence of lake surface water temperature, and Livingstone and Hari (2008) demonstrated that the lake surface temperatures of Fig. 17.3c also fluctuate coherently with river water temperatures throughout Switzerland. For lakes in the Tatra Mountains of Slovakia and Poland, Šporka et al. (2006) confirmed that surface water temperatures respond in a highly coherent fashion to regional climatic forcing, and that they decrease approximately linearly with increasing altitude. However, altitude can also affect the degree of coherence exhibited by lake surface water temperature. Although Livingstone et al. (2005a) showed that the surface water temperatures of lakes in the Swiss Alps respond coherently to regional climatic forcing over an altitude gradient exceeding 2,000 m, they also showed that at altitudes above ~2000 m a.s.l., the degree of short-term coherence in surface water temperature among the lakes decreases substantially, as does the strength of the relationship between lake surface water temperature and ambient air temperature. This is most likely to be the result either of a difference in the stratification regime at high altitudes, or of an increase in the influence of meltwater on the lake heat budget (Livingstone et al., 2005a).

Figure 17.3d illustrates the surface water temperature of Lake Balaton, Hungary, during the summer of 2000; i.e., during the same period of time as that covered by the measurements of surface water temperature in the Swiss lakes illustrated in Fig. 17.3c. A comparison of Fig. 17.3c with Fig. 17.3d makes it clear that the lake surface water temperatures in Switzerland and Hungary fluctuated coherently during this time (Livingstone and Padisák, 2007). This is perhaps surprising in view of the large distance separating the Swiss and Hungarian lakes (~750 km) and also in view of the fact that the lakes could hardly be more different in character, Lake Balaton being large (593 km^2) and located in the low-lying Carpathian Basin, and the Swiss lakes being smaller by a factor of $10^3 - 10^5$ ($0.0043 - 0.46$ km^2) and located in a mountainous environment, some at altitudes exceeding 2,000 m a.s.l. Livingstone and Padisák (2007) showed not only that the Swiss and Hungarian lake surface water temperatures fluctuated coherently, they also showed that these fluctuations mirrored fluctuations in the smoothed regional air temperature. This implies that an empirical model based solely on a spatially coherent regional air temperature should suffice to obtain a good estimate of short-term fluctuations in summer lake surface water temperature, even within a large, topographically heterogeneous region. Many studies have suggested methods of predicting lake surface water temperatures from air temperatures (e.g., McCombie, 1959; Webb, 1974; Shuter et al., 1983; Matuszek and Shuter, 1996; Livingstone et al., 1999). More recently, Kettle et al. (2004) have shown that the relationship between lake surface water temperature and air temperature can be improved by applying exponential smoothing to the air temperature. This approach, sometimes including other meteorological variables in addition to air temperature, has been shown to work well in a range of different lake districts, including topographically heterogeneous mountain regions (Kettle et al., 2004; Livingstone et al., 2005a; Šporka et al., 2006; Wilhelm et al., 2006; Livingstone and Padisák, 2007). In such empirical models, air temperature assumes a dual role. Firstly, as a causal variable it is involved explicitly in many of the processes that determine lake surface water temperature. Secondly, it is correlated with the other meteorological variables – viz. wind speed, cloud cover and relative humidity – that co-determine lake surface water temperature.

17.4 Long-Term Regional Coherence

Regional-scale coherence among lakes in the long term, here understood to refer to fluctuations on interannual to interdecadal time scales, was first brought to the attention of the limnological community by Magnuson et al. (1990). This study, based on a suite of lakes in Wisconsin, showed that long-term regional coherence was greatest for limnological variables that are directly influenced by climatic factors. Limnological variables that are influenced only indirectly by climate, or by common factors unrelated to climate (such as, for instance, by landscape position), exhibited less coherence. From the beginning therefore, the role of climate as the central factor responsible for long-term regional coherence in limnological variables was clear.

Further work on the Wisconsin lakes suggested that physical lake variables exhibit a higher degree of long-term regional coherence than chemical variables, which in turn exhibit a higher degree of coherence than biological variables (Kratz et al., 1998; Baines et al., 2000); subsequent work in other regions of the globe has confirmed this, for instance for the English Lake District (George et al., 2000). Here, examples are presented to illustrate the occurrence of long-term regional coherence in a range of physical, chemical and biological variables in selected European lakes and rivers.

17.4.1 Ice Cover and Water Temperature

17.4.1.1 Ice Phenology in Rivers and Lakes

Ice phenology – i.e., the timing of ice-on and ice-off – is treated in detail in Chapter 4 of this volume; here, we will simply summarise the information that is most relevant to the topic of long-term coherence. Finland and Sweden have a great number of lakes that consistently freeze over each winter, and for which long series of historical observations on ice phenology exist. In Finland, Palecki and Barry (1986) showed for 63 lakes that the timing of ice-on and ice-off were both highly correlated with air temperature over most of the southern part of the country. Livingstone (1999, 2000) showed that much of the coherent variability in the timing of ice-off, in Finland and elsewhere throughout the Northern Hemisphere, is closely related to the air temperature variability associated with the North Atlantic Oscillation (NAO) in winter, which is known to affect large areas not only of Europe, but of the Northern Hemisphere in general (e.g., Hurrell, 1995). Studies by Yoo and D'Odorico (2002), Blenckner et al. (2004) and Magnuson et al. (2004) confirm the importance of the NAO as a major contributor to coherence in ice phenology in northern Europe and elsewhere. Blenckner et al. (2004) found both the timing of ice-on and the timing of ice-off to be coherent among lakes in Finland and Sweden. In addition, they found both not only to be significantly related to the winter NAO, but also, even more strongly, to the regional atmospheric circulation, as represented by a set of regional circulation indices. A study of 196 Swedish lakes by Weyhenmeyer et al. (2004) also demonstrated the high degree of coherence in the timing of ice-off existing among lakes spanning a large latitudinal gradient (from 55°N to 68°N) in northern Europe. However, this study also highlighted the importance of including latitude, as a proxy for annual mean air temperature, in any analysis of the coherent response of lake ice phenology to climate forcing. This is not only for the obvious reason that ice-off occurs earlier in warmer, southerly regions than in colder, northerly regions, but, more importantly, because the timing of ice-off responds much more sensitively to air temperature in warmer regions than colder regions.

17.4.1.2 River and Stream Water Temperatures

An analysis of over 35 years of water temperature data from 25 sampling stations on rivers and streams throughout Switzerland, with catchment areas covering a large

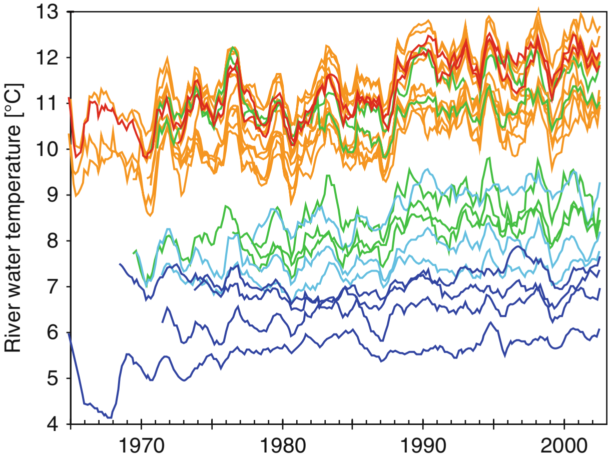

Fig. 17.4 River and stream water temperatures measured in Switzerland from 1965 to 2002 (12-month running means), colour-coded according to mean altitude h of catchment area: *red*, h < 500 m a.s.l.; *orange*, 500 m a.s.l. ≤ h < 1,000 m a.s.l.; *green*, 1,000 m a.s.l. ≤ h < 1500 m a.s.l.; *light blue*, 1,500 m a.s.l. ≤ h < 2,000 m a.s.l.; *dark blue*, h ≥ 2000 m a.s.l. After Hari et al. (2006)

range of mean altitudes (437−2,395 m a.s.l.), showed that river water temperatures in all parts of Switzerland exhibit spatially coherent interannual fluctuations (Fig. 17.4; Hari et al., 2006). The existence of this coherence, coupled with the general similarity shown by the temporal structure of the river water temperatures to that of the regional air temperature and the winter NAO (Hari et al., 2006), implies that the river water temperatures exhibit a common response to regional climatic forcing. The climatic factors responsible for the coherent, large step-change in river water temperature from 1987 to 1988 are currently the subject of investigation. The degree of regional coherence in all seasons was found to be particularly high in lower-lying areas and in the foothills of the Alps (i.e., for the higher temperatures shown in Fig. 17.4); however, coherence tends to decrease as the mean altitude of the catchment area of the sampling station increases (i.e., for the lower temperature curves of Fig. 17.4). The study also showed that spatial coherence in water temperature tends to be disproportionately low at sampling points influenced by either glaciers or hydro-electric power stations; this is because both meltwater from glaciers and deep water from reservoirs partially decouple the streams from regional climatic forcing. In the case of streams with partially glaciated catchment areas, inflowing meltwater diminishes the degree of coherence observed in summer and autumn. In their data set, Hari et al. (2006) showed that the three sampling sites with the lowest coherence each have over 14% glaciers and several hydro-power stations in their catchment areas. The sampling site with the lowest coherence has the highest percentage of glacier cover in its catchment area and the highest hydro-electric power production rate. Generalising from this study, it can be assumed that the water temperatures of rivers and streams are likely to respond coherently to climatic forcing over large areas, except when they are influenced by factors such as glacier meltwater, or when the hydrology of the system has undergone anthropogenic alteration.

17.4.1.3 Lake Surface Water Temperatures

Based on unique, 80-year-long time-series of measurements in eight lakes in Austria, Livingstone and Dokulil (2001) demonstrated that a high degree of spatial coherence in lake surface water temperature can exist over several hundred kilometres in all seasons. They showed this to result predominantly from the large-scale spatial coherence exhibited by air temperature across the region. In winter and spring, they showed that regional coherence in lake surface water temperature in their data is again ultimately traceable to the winter NAO, whereas in summer the processes responsible are less large-scale in nature. The influence of the winter NAO on the surface and near-surface lake temperatures of lakes throughout Europe has been confirmed in many studies (e.g., George et al., 2000; Straile and Adrian, 2000; Gerten and Adrian, 2001; Straile et al., 2003a,b; Weyhenmeyer, 2004; Blenckner et al., 2007). Despite the strong influence of the NAO on the winter climate, the occurrence of periods of ice cover of various duration was found by Livingstone and Dokulil (2001) to reduce the coherence in lake surface water temperature in winter. This was especially the case for higher-altitude lakes with long periods of ice cover. In spring and summer, the coherence in lake surface water temperature was found to be reinforced by a regional coherence in meteorological driving forces other than air temperature (e.g., wind speed in spring and high-altitude cloud cover in summer).

In northern Europe, Weyhenmeyer (2004) showed that the surface water temperatures of Sweden's largest lakes respond coherently to the winter NAO in March and May, and suggested that this coherent temperature response resulted in coherence being exhibited in various aspects of lake chemistry (see below). In the United Kingdom, surface temperature data from lakes in the English Lake District show a high degree of coherence in all months of the year, with regional coherence being exceptionally high in summer.

17.4.1.4 Deep-Water Lake Temperatures

Not surprisingly, hypolimnetic temperatures are generally much less coherent among lakes than are epilimnetic temperatures (Kratz et al., 1998; Benson et al., 2000). Nevertheless, in an analysis of coherence in the response of deep-water temperatures in 12 European lakes to climate forcing, Dokulil et al. (2006) showed regional coherence between lakes to be significant in the majority of cases, with the proportion of shared variance (r^2) being as much as 89% between Lake Constance (Austria/Switzerland/Germany) and Walensee (Switzerland), 79% between Attersee and Hallstättersee (both in the Salzkammergut region of central Austria) and 44% between Vänern and Vättern (both in southern Sweden). However, some lakes deviated somewhat from the regional norm: the deep-water temperature of Traunsee, for instance, one of the Austrian Salzkammergut lakes, was more closely correlated with the deep-water temperature of very distant lakes, such as Lake Constance and Lake Zurich (300–400 km to the west), and even Vänern (1,250 km to the north), than with that of any of its immediate neighbours. In contrast to lake surface water temperature, which can usually be upscaled without too much problem, one must therefore be extremely cautious in generalising from the behaviour of the deep-water temperature of one lake to that of its immediate neighbours. Coherence in the deep-water temperatures of four deep Swiss lakes was investigated by Livingstone (1993, 1997), who showed that, although these temperatures did exhibit a degree of regional coherence, this was greatly affected by the individual characteristics of the lakes. Because deep-water temperatures are influenced primarily by the meteorological conditions prevailing during spring turnover, deep-water temperatures tend to reflect the conditions pertaining during spring rather than integrating the annual climate. This implies that deep-water temperatures in western Europe also tend to be dominated by the NAO (e.g., Straile and Adrian, 2000; Gerten and Adrian, 2001; Straile et al., 2003a, b), although the relationship is more complex than that between the NAO and lake surface water temperature. During summer stratification, deep-water lake temperatures tend to increase gradually at an approximately constant rate as a result of the slow downward transport of heat from the lower metalimnion into the hypolimnion. During winter and spring in open-water lakes, heat loss to the atmosphere at the lake surface coupled with deep, penetrative mixing redistributes the heat in the lake, resulting in a rapid decrease in deep-water temperature. However, during warm winters, stratification can persist from autumn through to spring, inhibiting turbulent mixing and allowing the deep-water temperature to continue its gradual increase without interruption. This eventually results in an irregular sawtooth pattern in the deep-water time-series (Livingstone, 1993, 1997). Although a sawtooth pattern occurs in the deep-water temperature of very many lakes, the exact form of this pattern is critically dependent on individual lake characteristics such as lake depth, exposure to wind and degree of meromixis. Lakes within the same region that are subject to essentially the same climatic forcing can therefore exhibit completely different sawtooth patterns (Livingstone, 1993, 1997). Thus, although a high degree of spatial coherence may be present in the meteorological forcing variables and in lake surface water temperatures, this will not necessarily be the case for deep-water temperatures. Further information on the impacts of climatic variations and climate change on the thermal characteristics of lakes in Europe is given in Chapter 6 of this volume.

17.4.2 River Discharge Rate and Lake Level

In contrast to most limnological variables, hydrological variables, such as river discharge rate and lake level, are governed to a much greater extent by precipitation rather than heat balance. The spatial heterogeneity of precipitation is notoriously large in comparison to that of other meteorological variables; it is therefore interesting to note that hydrological variables – which result essentially from the integration

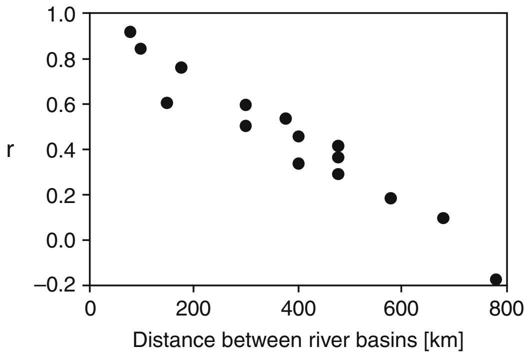

Fig. 17.5 Spatial coherence in monthly mean discharge rates of six rivers in Finland. The plot shows the correlation coefficient (r) between monthly mean discharge rates as a function of the distance between the individual river drainage basins. The correlations are based on data for 1961–2002 in the rivers Kokemäenjoki, Aurajoki, Tornionjoki and Kiiminkijoki, and for 1971–2002 and 1984–2002 in the rivers Kalajoki and Pyhäjoki

of precipitation within river drainage basins – can also exhibit a high degree of spatial coherence. A good example of this is provided by Fig. 17.5, which illustrates the spatial coherence found in the monthly mean discharge rates of six rivers in Finland. The geographical locations of the drainage basins of the rivers range from southern boreal to subarctic. Drainage basins separated by less than ~200 km were found to exhibit an extremely high degree of coherence in their monthly mean discharge rates, but the degree of coherence fell approximately linearly with increasing distance to reach zero at a distance of ~700 km.

In flat landscapes, rates of discharge from lakes are often limited by the low gradients of the outflowing rivers. In such situations, coherent variations in river discharge rates can result in large, coherent variations in lake water level. The example in Fig. 17.6 illustrates the substantial long-term coherence in water level that exists among several large lakes in north–eastern Europe. Since these lakes tend to have extensive shallow littoral areas, vertical variations in lake water level result in large variations in lake volume and in the relative position of the shoreline. In north-eastern Europe, coherent fluctuations in lake level can therefore be an important factor influencing lake ecosystems (Nõges et al., 2003; Nõges and Nõges, 2004). In Lake Võrtsjärv, for example, with a surface area of 270 km^2 and a mean depth of 2.8 m, the lake level fluctuates seasonally by 1.4 m on average; i.e., by half the mean depth of the lake. The absolute range of these fluctuations (3.2 m) even exceeds the lake's mean depth. Between the recorded minimum and maximum water levels, the lake surface area varies by a factor of 1.4, the mean depth by a factor of 2.5, and the volume by a factor of 3.5. The sensitivity of large lake ecosystems to changes in water level decreases with increasing mean depth. The seasonal and inter-annual ranges of water level fluctuations in Lake Peipsi are of the same magnitude as those found in Lake Võrtsjärv, but since Lake Peipsi has a much greater mean depth (7.1 m), water level variations affect this lake much less than they do Lake Võrtsjärv.

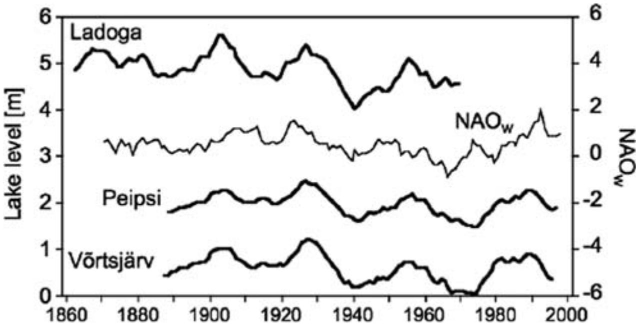

Fig. 17.6 Spatial coherence in annual mean water level of three large lakes in north–eastern Europe. Water level data for Lake Võrtsjärv (Estonia) and Lake Peipsi (Estonia/Russia) were obtained from the Estonian Institute for Meteorology and Hydrology; data for Lake Ladoga (Russia) were taken from Jaani (1973). All series were smoothed with a 7-year moving average. The time-series of the North Atlantic Oscillation index (Hurrell, 1995) is shown for comparison

Winter precipitation in northern Europe is known to be influenced significantly by the winter NAO, with precipitation being above average during a positive phase of the NAO and below average during a negative phase (Hurrell, 1995). Thus it is likely that spatial coherence in river discharge and lake level in this region may ultimately be related to the large-scale influence of the winter NAO. Various studies suggest that the NAO is indeed the driving force behind much of the variability in stream discharge in various regions of Europe. In northern and western Europe, above/below average values of river discharge tend to be associated with positive/negative phases of the winter NAO (e.g., Kiely, 1999; Hänninen et al., 2000; Bradley and Ormerod, 2001). In southern Europe, the situation is the reverse: coherent variations in the discharge rates of the Rhône and the Ebro appear to reflect variations in the winter NAO, with high rates of discharge being associated with the negative phase of the NAO, and low rates of discharge with the positive phase (Lloret et al., 2001). With respect to lake level variations, Rodionov (1994) demonstrated that the water level of the Caspian Sea was related in a more complex way to the NAO via its effects on precipitation rates in the Volga basin. From Fig. 17.6, it can be seen that longer-term coherent fluctuations in the water levels of Lake Ladoga, Lake Peipsi and Lake Võrtsjärv also appear to reflect the behaviour of the winter NAO. A similar behaviour has been documented for lakes in northern Germany (Behrendt and Stellmacher, 1987) and Russia (Doganovsky and Myakisheva, 2000).

17.4.3 Lake Chemistry

Investigations of regional coherence in the chemical properties of lakes have focused mainly on the effects of atmospherically transported pollution, which have been dealt with in great detail in the literature (e.g., Battarbee et al., 2005) and which will thus not be considered further here. Somewhat less attention has been paid to regional coherence in lake nutrient concentrations (e.g., George et al., 2004a,b; Weyhenmeyer, 2004), which will therefore form the focus of this section. Some nutrient studies have been based on raw time-series, whilst in others, the data were detrended to remove the effects of progressive alterations in trophic status brought about by large-scale changes in land use or by the simultaneous introduction of measures to counter eutrophication.

The most comprehensive recent investigation of long-term regional coherence in lake chemistry was undertaken by Weyhenmeyer (2004), who studied NAO-related synchrony in the behaviour of 13 chemical variables at 16 sites in Sweden's largest lakes, Vänern, Vättern and Mälaren. She found that coherent relationships between the winter NAO and water chemistry among lake sites were restricted to variables closely linked to surface water temperature, and that the strongest coherent response to climate forcing was exhibited by reactive silica and pH in May, after the ice on the lakes had thawed. In a comparable study of 28 boreal lakes located within a circle of radius 10 km in southern Finland, Järvinen et al. (2002) showed interannual coherence to be high for conductivity, calcium and alkalinity (in addition to water temperature). Interannual fluctuations in local weather explained part of the interannual fluctuations in lake chemistry, but pairs of lakes that were directly connected via a surface stream were found to exhibit a higher coherence than pairs of lakes that were not, implying that the presence or absence of hydrological connectivity can affect the interannual coherence of chemical variables. The high coherence found by Järvinen et al. (2002) for calcium agrees with similar results obtained by Baines et al. (2000) for the lakes of northern Wisconsin. Presumably because the winter ice cover effectively insulated the Finnish lakes from climatic forcing, the water chemistry under the ice in March appeared to be associated with the weather conditions prevailing during the previous autumn rather than winter.

Other chemical variables that exhibit a degree of coherence in their behaviour that appears to be linked to the winter NAO include total organic carbon (TOC), oxygen and soluble reactive phosphorus. In a study of the influence of climate and land use on rivers in Finland, Arvola et al. (2004) found TOC loads in spring to respond coherently to the winter NAO in rivers in northern Finland, but not in southern Finland. Investigations into deep-water oxygen concentrations in Swiss lakes have revealed a connection to the suppression of circulation during warm winters (Livingstone, 1997; Straile et al., 2003a), which can also be assumed to be linked to the winter NAO. In the case of phosphorus, coherence seems to be much lower, as does its association with the NAO (e.g., George et al., 2004b). However, there is some evidence suggesting the existence of regionally coherent fluctuations in soluble reactive phosphorus in lakes. Blelham Tarn in the English Lake District is quite a productive lake surrounded by agricultural land, but there has been no recent increase in winter concentrations of soluble reactive phosphorus (Fig. 17.7a). Esthwaite Water, a neighbouring lake, has the added complication of a sewage treatment plant, so the progressive increase in the numbers of tourist visitors to the Lake District has resulted in increasing concentrations of this nutrient (Fig. 17.7b). The time-series of residuals generated by linear detrending (Fig. 17.7c) can be seen to be quite coherent, with 35% shared variance. The key factor underlying the coherence

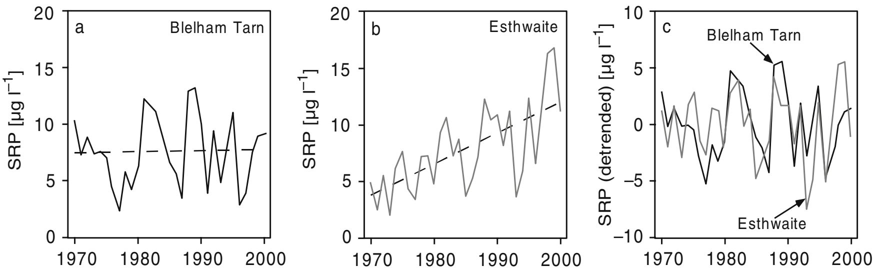

Fig. 17.7 The year-to-year variation in the winter concentration of soluble reactive phosphorus (SRP) in (**a**) Blelham Tarn and (**b**) Esthwaite Water, in the English Lake District. The *broken lines* show the linear regressions used to detrend the two time-series. (**c**) The year-to-year variation in the detrended SRP time-series for the two lakes

in soluble reactive phosphorus in this case is the interannual variation in winter rainfall. The surface runoff following a heavy rainfall washes soluble reactive phosphorus into lakes from their (unfrozen) catchment areas, which, in the case of lakes with short retention times, results in higher concentrations of this nutrient being recorded (George et al., 2004b). Thus regional coherence in winter rainfall, possibly linked to the winter NAO, will result in regional coherence in winter concentrations of soluble reactive phosphorus in frost-free regions.

However, the chemical constituent that has attracted most attention in European lakes and rivers with regard to interannual coherence is nitrate. Nitrate concentrations in winter and early spring are known to exhibit strongly coherent interannual fluctuations related to winter climate (see Chapter 10 of this volume). In Fennoscandia, for instance, detrended nitrate concentrations in March show a significant positive correlation with the mean air temperature during the previous winter (George et al., 2004a; Weyhenmeyer, 2004). Reasons given for this include a reduction in nitrate release from frozen ground in the catchment area in cold winters (Weyhenmeyer, 2004) and the earlier melting of snow in warm winters (George et al., 2004a). By contrast, in western and central Europe, winter and spring nitrate concentrations in lakes and streams appear to vary inversely with mean winter air temperatures (George, 2000; Monteith et al., 2000; George et al., 2004a,b; Davies et al., 2005; Jankowski et al., 2005). The most likely explanation for this widespread response would seem to be the effect that the severity of the winter has on the terrestrial assimilation of nitrate in the catchment area. The higher denitrification rates of soil bacteria that occur in milder winters, in combination with higher nitrate leaching due to frost damage of biological material in cold winters, may be responsible for the negative correlation of nitrate concentrations with winter air temperature in the UK and Switzerland (Monteith et al., 2000; George et al., 2004a,b; Davies et al., 2005; also see Chapter 10 of this volume). Nitrate leaching is enhanced during cold winters when a greater duration and intensity of soil freezing enhances biocidal effects, releasing more nitrogen for mineralisation, while low soil temperatures may also retard assimilation of nitrogen by soil biota. Not surprisingly, the high degree of coherence observed in variations in winter nitrate concentrations can again be traced back to the winter NAO (Monteith et al., 2000, George et al., 2000, 2004a,b). In Ireland, however, the pattern differs from that observed in the UK. Detrended nitrate concentrations in Lough Leane show no significant correlation with the winter NAO (Jennings et al., 2000). A recent analysis shows that interannual fluctuations in nitrate concentrations in Lough Leane and Muckross Lake are related to interannual fluctuations in the position of the Gulf Stream (Jennings and Allott, 2006). This relationship is driven by the effect the position of the Gulf Stream appears to have on moisture levels in catchment soils during the previous summer. Low soil moisture levels inhibit the uptake of nitrogen by plants and soil microbes, leading to higher soil nitrate availability in the autumn and winter, when wetter conditions re-occur (Jennings and Allott, 2006).

Various factors can influence the long-term regional coherence of nitrate concentrations in rivers and lakes. Local geomorphological lake characteristics, such as lake size, the ratio of surface area to mean lake depth or the ratio of lake surface area to catchment area, are likely to modify the effect of climatic forcing on the regional coherence of nitrate. In the central perialpine region, for example, nitrate concentrations in smaller lakes appear to respond more sensitively to climatic forcing than in larger lakes (Jankowski et al., 2005). In addition, Blenckner et al. (2007) suggest that the ratio of surface area to mean lake depth may significantly affect the strength of the climate signal. The strength of the regionally coherent response of nitrate concentration to large-scale climatic forcing factors may also undergo a gradual long-term change that is driven by external forces, such as changes in nitrogen deposition, as in the case of lakes in Sweden (Weyhenmeyer et al., 2007), or the use of fertilisers in lake catchment areas, as in the English Lake District (George et al., 2000). In the English Lake District there appears to be a marked seasonal variability in the regional coherence of nitrate related to climate forcing: although regional coherence in mean nitrate concentration in the Lake District is high in all months, it is highest during winter and spring (Fig. 17.8). This is predominantly a result of the effect of interannual variations in the weather in winter and spring on the leaching of fertilisers applied to the surrounding land. George et al. (2004b) have shown that the concentrations of nitrate reaching the lakes are much lower in

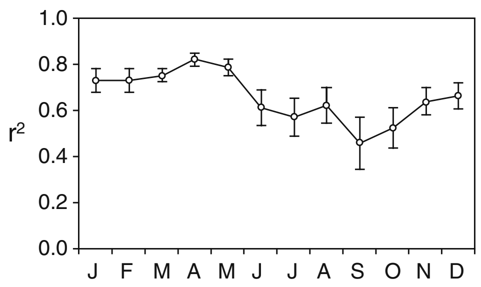

Fig. 17.8 Seasonal variability of the pairwise coefficient of determination (r^2) of the linearly detrended mean nitrate concentrations in Blelham Tarn, Esthwaite Water and the North and South Basins of Windermere in the English Lake District (means and standard errors). Calculations based on data from 1960–2000

mild winters when there is an enhanced uptake of nitrate in the surrounding land. In summer, the concentrations of nitrate in the lakes are reduced by biological uptake within both the lakes and their catchments, but the residual concentrations present are still quite high, so the effect of coherent fluctuations in nitrate concentrations in winter and spring effectively persists into summer.

17.4.4 Lake Biology, Population Dynamics and Food Web Interactions

The degree of coherence observed in most biological time-series is much lower than in physical and chemical time-series because of the multiplicity of confounding factors involved. Nevertheless, coherence studies of biological variables are of value in highlighting common biological responses in lakes exposed to comparable large-scale driving forces.

Biological populations in seemingly isolated lacustrine ecosystems may exhibit coherent behaviour if aspects of those elements of the physical environment that regulate their temporal dynamics also exhibit a degree of coherence. Many biological processes in lakes are strongly influenced by physical processes, but the impact of these processes varies from season to season and from lake to lake. In temperate lakes, the physical conditions that prevail during winter and spring differ strongly from those required for optimum growth. Water temperatures are low, resulting in low growth rates of zooplankton, benthos and fish. The low solar elevation, sometimes combined with ice and snow cover, results in low light intensities in the water column and hence in low phytoplankton growth rates. The vigorous turbulent mixing that normally occurs either in winter and early spring or (in the case of ice-covered lakes) after the ice has thawed also affects the availability of light to the phytoplankton. In spring, lake phytoplankton communities are typically dominated by fast-growing species that are adapted to steep light and temperature gradients, and are thus particularly responsive to the physical conditions that occur at this time of year (Adrian et al., 2006). By contrast, physical conditions during summer (high temperatures and high light availability) are much closer to those required for optimum growth, so that population dynamics are increasingly influenced by other factors, such as nutrient limitation or biotic interactions (Sommer et al., 1986; Straile, 2005). Moreover, during summer, slowly growing zooplankton species with longer and more complex life cycles are often quite abundant. Because these species can be affected by multiple warming events in the course of a year, their response to environmental changes tends to be more species-specific (Adrian et al., 2006; Blenckner et al., 2007). Hence, in temperate lakes biological variables are more likely to show coherence during winter and spring than during the rest of the year. In the following we will discuss examples of coherence in phytoplankton (see also Chapter 14 of this volume), zooplankton, and fish.

Phytoplankton growth during winter and early spring is severely light limited (Tulonen et al., 1994). Intrinsically low incident light levels can be further reduced within the water column by ice and snow cover (Adrian et al., 1999; Weyhenmeyer et al., 1999), while vigorous, penetrative turbulent mixing will reduce light availability for phytoplankton in deep lakes (Peeters et al., 2007). Consequently, the timing of the phytoplankton spring bloom in ice-covered lakes might also exhibit coherence if the timing of ice-off shows a coherent response to climatic forcing. The timing of the spring phytoplankton bloom in Müggelsee, Germany, and Erken, Sweden, has been shown to depend on the timing of ice-off and on the NAO (Adrian et al., 1999; Weyhenmeyer et al., 1999; Gerten and Adrian, 2000), suggesting the existence of coherence in phytoplankton phenology over large distances. However, no coherence in spring phytoplankton bloom dynamics was observed between Müggelsee, in northern Germany, and Lake Constance, in southern Germany (Straile and Adrian, 2000), as the mechanisms controlling the onset of the spring bloom – i.e., the timing of ice-off (Müggelsee) and the timing of the onset of stratification (Lake Constance) – differed between the lakes.

Studies on Sweden's largest lakes (Vänern, Vättern and Mälaren), and in much smaller Lake Erken, have shown a tendency towards an earlier spring phytoplankton bloom and earlier nutrient depletion (of up to one month) in all lakes as winter air temperatures have increased (Weyhenmeyer et al., 1999; Weyhenmeyer, 2001). In Sweden's largest lakes the total phytoplankton biomass from May to October showed no coherent change, but the spring and summer biomass of phytoplankton groups such as cyanobacteria and chlorophytes increased coherently among the lakes as air temperatures increased. After the especially mild winter of 1988/1989, an increase in cyanobacterial biomass was found throughout Europe (Weyhenmeyer et al., 2002). Considering that some of the cyanobacterial species that dominate in these lakes in summer can be toxic, the large-scale, coherent effect of warmer winters on phytoplankton is potentially far-reaching.

Changes in the structure of the phytoplankton community can also be coherent, as indicated by a multivariate study of phytoplankton species composition in five, mainly deep, perialpine lakes; viz. Lower Lake Zurich, Upper Lake Zurich, the Lake of Walenstadt (Walensee), Lake Geneva and Lake Constance (Anneville et al., 2004, 2005). Trophic status and local environmental conditions are different among the lakes, and the phytoplankton assemblages found in the lakes are also different. Nevertheless, the patterns of occurrence of some phytoplankton assemblages were found to be strongly synchronous on annual to decadal time-scales. This long-term regional coherence sometimes involves different phytoplankton assemblages. Although the composition of the phytoplankton in these lakes was strongly influenced by the long-term reduction in phosphorus loading, changes in seasonal meteorological forcing were also found to have induced synchronous changes in the phytoplankton assemblages. These results provide evidence that synchronous long-term changes in the phytoplankton communities of geographically distant lakes can occur despite there having been no synchrony in overall biomass changes.

While the phytoplankton coherence observed in these lakes was due mainly to a coherent decline in nutrient concentrations, statistical analysis also suggested the existence of a link between community structure and the winter NAO. Surprisingly, even summer phytoplankton communities exhibited a coherent response to climatic forcing in these lakes, but the mechanisms responsible for this are not yet clear

(Anneville et al., 2005). In Esthwaite Water and Blelham Tarn, two productive lakes in the English Lake District, the key factor responsible for the observed coherence in the summer abundance of the zooplankton proved easier to identify (George et al., 2000). Here, it was the mixing events that entrained nutrients from the deep water and stimulated the growth of edible algae that were found to be critical.

The rates of physiological processes in ectothermal organisms are strongly influenced by ambient temperatures. Consequently, the growth and reproduction rates of zooplankton species will be limited by low temperatures during winter and spring. As coherence in the timing of the spring increase in water temperature is high on regional and even supra-regional scales, we might expect coherence in zooplankton growth on similar spatial scales to occur during spring. Despite great limnological differences between Müggelsee and Lake Constance, the growth of a key herbivorous zooplankter, *Daphnia*, was indeed found to be highly coherent in these two lakes, and in both cases was related to the winter NAO (Straile and Adrian, 2000). Furthermore, the timing of the clear-water phase – i.e., the timing of maximum water transparency during early summer due to *Daphnia* grazing – was also found to be coherent (Straile and Adrian, 2000). Subsequent studies in a number of European lakes demonstrated the existence of similar links between the winter NAO and the spring biomass of *Daphnia*, the timing of the *Daphnia* spring peak and the timing of the clear-water phase: for example, the observations by Scheffer et al. (2001) in a suite of Dutch lakes, by Straile (2002) in central Europe, by Anneville et al. (2002) in Lake Geneva and by Wagner and Benndorf (2007) in Lake Bautzen in Germany.

There have been very few coherence studies on aquatic invertebrates and fish. However, George (2000) and Elliott et al. (2000) have shown that the times of emergence of alder flies and sea trout fry in the English Lake District are correlated with the NAO. The most comprehensive study of coherence in the spawning time of fish is that of Nõges and Järvet (2005) in Estonia. Their analysis was based on observations conducted over a period of 40 years (1951–1990) at 148 stations in the region of Estonia which includes the two largest lakes in the country; viz. Võrtsjärv and Peipsi. In both lakes, a coherent, long-term shift in the timing of spawning of bream (*Abramis brama*) was found (~10 days earlier over the 40-year period) that tended to compensate for the long-term increase in water temperature. By contrast, the timing of the spawning of roach (*Rutilus rutilus*) was approximately constant regardless of the long-term increase in water temperature. However, interannual fluctuations in the spawning date of roach tended to occur coherently between the lakes, although there was no long-term shift. No evidence was found of coherent interannual fluctuations in the spawning date of bream. As the earlier spawner, roach is presumably influenced more strongly by the more coherent early spring changes in lakes, while the potential for coherence in interannual fluctuations disappears by the time of spawning of bream (on average 16–17 days later than roach).

Thus ecological responses to fluctuations in climate can occur coherently over large spatial scales provided that physical forcing is coherent and also rate-limiting. Given these prerequisites, coherent behaviour can be observed in the phenology of life-cycle events and in the interactions that regulate the dynamics of food webs in many aquatic systems.

17.5 Long-Term Supra-Regional Coherence

In the above sections, we have presented examples that demonstrate the existence of spatial coherence in limnological variables on regional scales; i.e., from about ten kilometres up to several hundred kilometres. However, climate-driven spatial coherence exists on even larger scales. For example, Benson et al. (2000) compared time-series of temperature from lakes in four lake districts in central North America that are separated by up to 1,300 km, and found strong coherence in near-surface temperatures and epilimnetic temperatures (but much weaker coherence in hypolimnetic temperatures). In a study of coherence in various lake variables within and among regions on the Canadian Boreal Shield, Arnott et al. (2003) also found significant supra-regional coherence in near-surface temperatures, but much more heterogeneity in pH and in the biomass and richness of phytoplankton and zooplankton. In Europe, a recent extensive study (Blenckner et al., 2007) explored the spatial coherence in the response of a suite of physical, chemical and biological variables to climatic forcing on spatial scales exceeding ~1000 km. Because of the well-established influence of the NAO on many European lakes, they used a meta-analysis approach to summarise the observed correlations between the selected limnological variables and indices of the winter NAO. Results were collated from 7 lakes in northern Europe (Estonia, Finland and Sweden), 7 lakes in central Europe (Austria, Germany and Switzerland) and 4 lakes in western Europe (the United Kingdom and Ireland). Not unexpectedly, they found the strongest response to the winter NAO was exhibited by lake water temperatures, especially by the surface water temperature. Again not unexpectedly in view of previous work on individual lakes, winter nitrate concentrations also showed a strongly coherent response to the winter NAO, with concentrations tending to be high in positive NAO years and low in negative NAO years. Notable exceptions were the lakes of the English Lake District, where nitrate concentrations were lower in positive NAO years, when more nitrate was assimilated in the soils of the surrounding catchments. This difference had already been noted in an earlier study of the correlations observed between the detrended winter nitrate concentrations in Finland and the UK and the NAO (George et al., 2004a). In Finland, the correlation with the NAO was negative and clearly related to the timing of snow melt. In the UK, the correlation was negative and closely correlated with the winter air temperature. Blenckner et al. (2007) also noted a tendency for the concentrations of soluble reactive phosphorus and soluble reactive silicate to be negatively correlated with the winter NAO. The authors suggested that this behaviour may be related to large-scale coherent interannual fluctuations in the phenology of the spring phytoplankton bloom, but the response did not appear to depend on the trophic status of the lakes. Interannual fluctuations in winter and spring concentrations of total phosphorus were also related to the winter NAO, with low values in positive NAO years and high values in negative NAO years. However, the winter and spring phytoplankton biomass showed no coherent relationship to the winter NAO. By contrast, the abundance of daphnids in winter and spring was found to fluctuate coherently with the winter NAO, although lake-to-lake variability was high. Calanoid copepods showed no overall coherent relationship to the winter NAO, but the summer abundance of cyclopoid copepods was significantly higher in positive NAO years than in negative NAO years, possibly as a result of coherent fluctuations in the timing of their emergence from resting stages.

17.6 Conclusions

In this chapter we have summarised the results of studies designed to investigate the degree of short-term and long-term coherence apparent on regional and supra-regional scales among lakes and rivers in Europe. These results allow us to answer the original question posed in Fig. 17.2; viz. ‘Are local effects strong enough to destroy the coherence imparted by the climate?’ In general the answer is quite clearly no. Although the meteorological variables that ultimately govern much of the dynamic behaviour of lakes and rivers act locally, these forcing variables often exhibit a high degree of coherence over a range of spatial and temporal scales. Coherence in the behaviour of the driving meteorological variables will almost invariably result in a degree of coherence in the response of water bodies forced by these variables. Local effects, both external and internal to the water body, add noise to the climate signal, but the magnitude of the observed coherence present allows us to reject the hypothesis that surface water bodies respond merely as idiosyncratic individuals to climate forcing. From the point of view of climate impact research, surface water bodies can be regarded conceptually as local samples of a climatically-driven continuum.

For lakes, spatial coherence is, not surprisingly, highest for physical variables measured at the lake surface. Short-term coherence in lakes is essentially confined to water temperatures in the uppermost part of the water column. On interannual to interdecadal time-scales, the timing of ice-out, the temperatures of rivers, streams and lake surfaces and the discharge rates of streams are all directly influenced by regional climatic forcing. They consequently exhibit a high degree of coherence that is linked to the regional-scale spatial homogeneity of the relevant meteorological driving variables. Air temperature, which exhibits the highest degree of spatial homogeneity, is responsible for much, but not all, of the observed coherence. Since air temperatures in much of Europe in winter and spring are strongly influenced by the climate prevailing over the North Atlantic, it is not surprising that physical lake surface and river variables at these times of year are strongly linked to the NAO. As the NAO affects a number of meteorological driving variables, its influence on lakes is typically greater than that of any single variable.

Water temperatures below the thermocline are invariably less coherent from lake to lake, since they are strongly influenced by the mixing characteristics of the individual lakes. However, because the physical (and chemical) characteristics of the hypolimnion are determined to a large degree during spring turnover, the weather conditions prevailing during turnover often leave their signature in the deeper water for a considerable length of time. As the climate in northern and western Europe at this time of year is often strongly influenced by the NAO, this implies that a coherent NAO signal can persist for some time in the deep water.

Coherent behaviour in nutrient concentrations appears to be strongest in the case of nitrate, which often exhibits an interannual variability linked to the NAO. Concentrations of soluble reactive phosphorus appear to be influenced by the climate to a much lesser extent, but may exhibit some decadal-scale coherence related to anthropogenic effects. Coherence in biological variables is typically very much weaker, but coherent patterns are frequently reported in derived variables that reflect the timing of biological events; e.g., the timing of the spring bloom and the clear-water phase.

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Chapter 18
The Impact of Climate Change on Lakes in Northern Europe

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18.1 Introduction

In Northern Europe, most lakes are characterized by extended periods of winter ice cover, high spring inflow from snow melt and brown water produced by the transport of dissolved organic carbon (DOC) from the surrounding catchments. In this chapter, the potential impact of climate change on the dynamics of these lakes is addressed by: (i) Describing the historical responses of the lakes to changes in the weather. (ii) Summarizing the results of modelling studies that quantify the impact of future changes in the climate on the lakes and the surrounding catchments. Many existing water quality problems could well be exacerbated by the effects of climatic change. It is therefore important to assess the holistic responses of the individual lakes to the combined effects of local changes in the catchment and regional changes in the weather (Hall et al., 1999; Anderson et al., 2005). Overall, the response of individual lakes to climate change can be very different (Blenckner et al., 2004). For example, mountain lake catchments are affected differently from those at lower altitudes. In addition, the landscape position of a particular lake influences hydrological flow regime (Kratz et al., 1997). Furthermore, the response of lakes to climatic variation is also modified by physical lake features such as morphometry and water clarity which, in turn, is also affected by the concentration of the dissolved organic carbon (see for example Fee et al., 1996). Also, the alignment of the lake in relation to the main wind direction is important for the timing of the ice break-up and mixing regime. Even, the environmental changes experienced by the lake in the past can affect the magnitude of the response to climatic variation. Lakes in a recovery phase from eutrophication, acidification, toxic components or any other strong human disturbance, might respond differently to climatic variability and change owing to their specific history and food web structure.

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In this chapter we will: (a) Introduce the Northern landscape and its climate. (b) Describe both the coherent and individual responses of the lakes to the observed variations in the climate and the changes projected for 2070–2100. (c) Summarize these results in a form that is more accessible to potential end-users.

18.1.1 Lakes in Northern Europe

Lakes form an important component of the Northern European landscape (Fig. 18.1). The area of the landsurface covered by lakes in this region is very much greater than in other parts of Europe and ranges from 4.4% in Estonia, to 8.7% in Finland and 9% in Sweden (EEA database, www.eea.europa.eu). The numbers

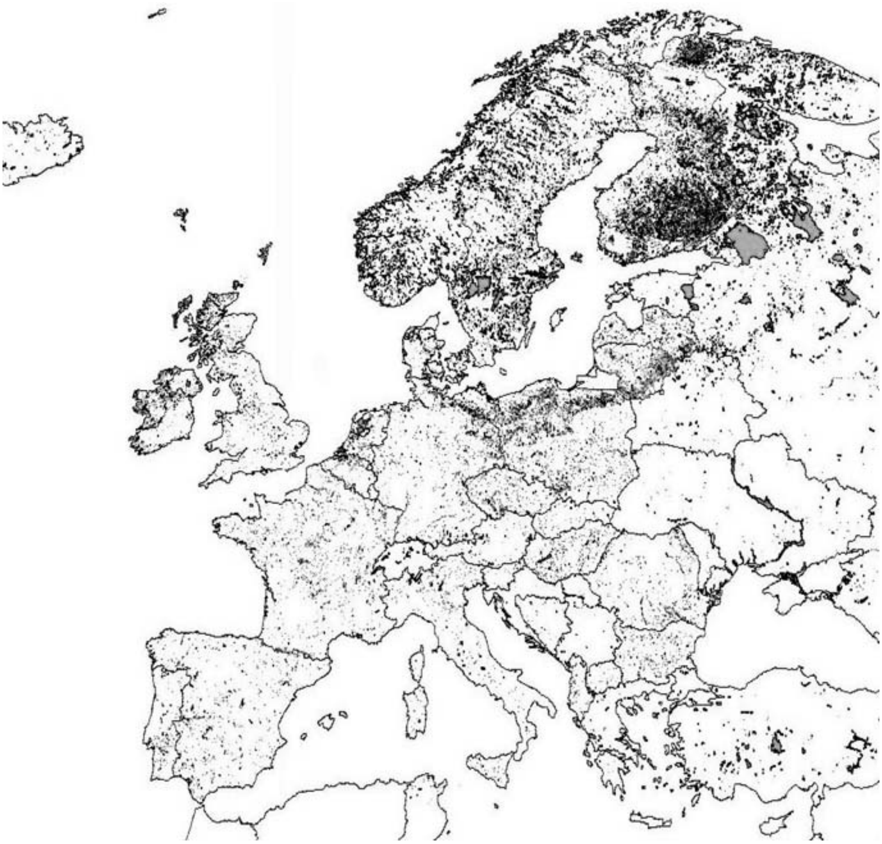


Fig. 18.1 Map showing the high concentration of lakes (*dots*) in Northern Europe. The number of lakes is based on Article 3 submissions of the Water Framework Directive and automatic detection of surface water using Landsat 7 Imagery Erika Rimaviciute and Alfred de Jager, European Commission 2006

of lakes in Finland and Sweden are also very large. In Finland, there are 187, 888 lakes with a surface area >0.01 km^2 (Raatikainen and Kuusisto, 1990) and the corresponding figure for Sweden is 65,000 (http://www.smhi.se/). These lakes are all relative young and were formed after the last ice age, ca. 10,000 years ago. Small (<0.1 km^2) and shallow (mean depth <5 m) lakes are most numerous and many of the shallow lakes have poor light penetration since they contain high concentrations of coloured humic compounds. The largest lakes in the region are Vänern and Vättern in Sweden, Saimaa, Inari, and Päijänne in Finland and Peipsi and Võrtsjärv in Estonia. In trophic terms, the lakes cover a wide range of nutrient concentrations. The most eutrophic lakes are those situated in southern Sweden, Estonia and the coastal area of Southern Finland. The lakes situated further to the north are usually oligotrophic and a high proportion are located in forest areas contain high concentrations of coloured humic substances. Many of these lakes are naturally acidic but some lakes located in southern Fenno-Scandia are still impacted by anthropogenic acidification. In Sweden, many of these acid lakes are treated with lime to increase the pH (Henrikson and Brodin, 1995). In the whole region, most of the lakes are covered with ice throughout the winter but there are large spatial variations (Weyhenmeyer et al., 2004 and Chapter 4, this volume) and some large, deep lakes in southern Sweden and Norway can be ice-free during winter.

18.1.2 Climate in Northern Europe

The climate in Northern Europe ranges from temperate to sub-arctic (Chapter 2, this volume). In the areas nearest to the CLIME sites, the annual mean temperature recorded during the IPCC reference period (1961–1990) varied from 3.4 to 6°C with an annual precipitation variation of 500–632 mm (Table 18.1). Model scenarios of the future change in the climate (Chapter 2) suggest that the annual temperature could be 2–6°C higher by 2071–2100 with the most pronounced increases (4–5°C) being projected for the winter (December–February). Some scenarios also suggest that there will be a 3–4°C increase in the summer (June–August) temperature with the most pronounced increases projected for Estonia. The projected changes in the precipitation are more variable and range from a minimum of -10 mm $year^{-1}$in Estonia in summer to a maximum of +50 mm $year^{-1}$ in Southern Norway in winter (see Chapter 2 for more details). All the scenarios suggest that there will be a drastic decrease in the winter snow cover.

18.2 Physical Responses

18.2.1 Historical Changes

A key feature of the Northern European lakes is their winter ice and snow cover which influences water temperature, light penetration, the supply of nutrients and

Table 18.1 Climate statistics from stations close to three of the Northern CLIME sites. The annual mean of the air temperature and precipitation of the observed climate, obtained from the Climate Research Unit in the UK and of the three, by two different climate circulation models (Chapter 2), simulated periods, i.e. control (1961–1990) and the two future (2071–2100) emission scenarios A2 and B2

	Erken, Sweden	Pääjärvi, Finland	Peipsi, Estonia
Air temperature (°C):			
Observed (1961–1990)	6.0	3.4	4.9
Control (1961–1990)	6.0	4.0	5.4
A2 (2071–2100)	10.2	8.5	10.1
B2 (2071–2100)	9.2	7.4	8.8
Precipitation (mm):			
Observed (1961–1990)	500	600	632
Control (1961–1990)	635	785	766
A2 (2071–2100)	617	801	740
B2 (2071–2100)	599	783	700

the dynamics of the phytoplankton. For example, a series of mild winters with earlier ice break-up can lead to an earlier stratification and a shift in the composition of the phytoplankton from diatoms to cyanobacteria (Adrian et al., 1995). The meteorological factor that has the most pronounced effect on ice dynamics is the local air temperature (Palecki and Barry, 1986; Robertson et al., 1992; Vavrus et al., 1996; Livingstone, 1997). Warmer winters strongly advance the timing of the ice break-up (Adrian and Hintze, 2000; Yoo and D'Odorico, 2002). Ice cover data collated from lakes in the Northern Hemisphere over the last 150 years (Magnuson et al., 2000) show that these break-up dates have advanced by an average of 6.5 days per 100 years and the dates of first freezing delayed by an average of 5.6 days per 100 years. A number of investigators have now documented the effects that this change in the duration of ice-cover and the timing of freezing and thawing have had on the seasonal dynamics of the lakes (Salonen et al., 1984; Weyhenmeyer et al., 1999; Järvinen et al., 2002). In the last thirty years, there has been a dramatic change in the timing of ice break-up in many European lakes (Chapter 4, this volume) with the variation dictated by the geographic location of the lake. For some lakes in Northern Sweden and Finland the change in the timing of ice break-up over the last 30 years is almost zero, but it now close to 30 days for some lakes in Southern and Middle Sweden. The timing of the ice break-up of 50 lakes in Sweden and Finland were significantly negatively related to the North Atlantic Oscillation (NAO) winter index (Fig. 18.2). The most pronounced changes have been recorded in the south of Sweden where some lakes were totally ice free during the mild winters of 1990. Also in the large Swedish lakes Vänern, Vättern and Mälaren, the number of years without continuous ice-cover has increased in the 1990s (Weyhenmeyer, 2001). In northern Sweden (north of 62°N), the change is much less pronounced or totally absent

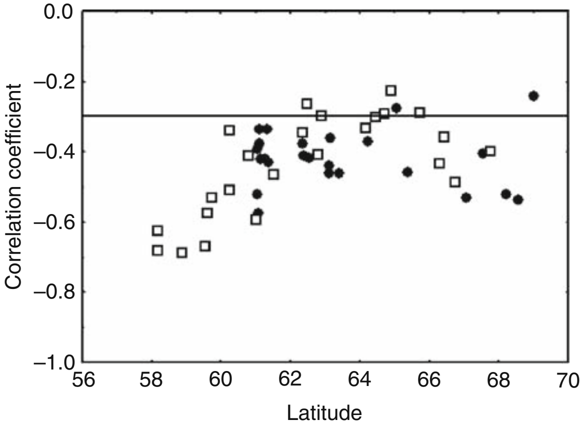


Fig. 18.2 The correlations between the ice break up dates at different latitudes in Sweden (□) and Finland (●) and the winter values of the North Atlantic Oscillation Index (NAOw). The *line* represents the 0.05 significance level. Modified from Blenckner et al. (2004)

(Blenckner et al., 2004). During the last 30 years there was a clear trend towards a higher variability in the ice break-up dates in the south of Sweden compared to the north, with a distinct change in the relationship again at around 62°N (Blenckner et al., 2004 and Chapter 4, this volume). This tendency was less pronounced for the timing of freezing and totally absent for the ice break-up dates of the Finnish lakes (Blenckner et al., 2004). A nonlinear relationship between air temperature and lake ice break-up, found in Swedish lakes, results in distinct differences in the way in which the recent change in the weather has influenced the timing of ice break in the colder and milder parts of Northern Europe (Weyhenmeyer et al., 2004).

In Chapter 6, this volume, Arvola et al. have described some of the factors influencing the long-term change in the surface and bottom temperatures of lakes located in Northern, Western and Central Europe. In Northern Europe, most of the lakes are frozen so records of the winter water temperatures are quite rare. An analysis of the surface temperatures variations recorded in 16 Northern lakes during the growing season (May–October) has, however, shown that these lakes have responded in a coherent way to the recent change in the weather (Weyhenmeyer et al., 2007, Fig. 18.3). This data was collated from 14 shallow lakes in Sweden, a lake in northern Germany and another lake in Estonia. Spring water temperatures throughout the region have shown a clear upward trend, a pattern partly explained by the increase in the spring air temperature and partly by the change in the winter water temperature (Nõges, 2004; Weyhenmeyer, 2004). The data collated from a number of CLIME sites and some other lakes demonstrate that there have been several abrupt shifts in their surface temperatures (ST) over recent decades (see Chapter 6 this volume). In Võrtsjärv and Lake Erken, an abrupt shift occurred in 1987 and the annual ST recorded during the previous 10 years was significantly ($p < 0.01$) lower (Nõges, T. and Blenckner, T., unpublished data). In contrast, long-term water temperature measurements acquired from eight large Finnish lakes between 1961 and 2000 did

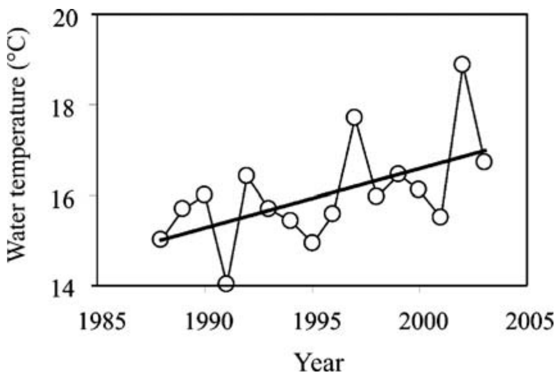


Fig. 18.3 The increase in the median 'open water' (May–October) surface temperatures in 16 European lakes. (modified from Weyhenmeyer et al., 2007)

not show any statistically significant warming (Korhonen, 2002). In some cases, clear trends were also recorded in the deep water temperature of the lakes. For example, the deep water temperatures of the two largest lakes in Sweden, Lake Vättern and Lake Vänern, show a long-term trend (1980–2003/2004) and the inter-annual variation in the winter NAO (NAO_W) explained 40–60% of the recorded variation in their deep water warming (Dokulil et al., 2006).

18.2.2 Future Projections

Comprehensive climate change experiments based on regional climate model (RCM) projections simulated a 1–2 month reduction in the lake ice season in Northern Europe for the period between 2070 and 2100. The largest changes were projected for lakes in southern and central Sweden, the southwest Baltic States and western Norway (Räisänen et al., 2001). Water temperature and ice phenology has also been modelled in five Finnish lakes (Elo et al., 1998). These model results also showed that the length of the ice-cover period decreased in all lakes, and there was an associated increase in the probability of intermittent ice-break and re-freezing (non-continuous ice-cover) during winters at the end of the 21st century. The ice-covered period of Lake Pääjärvi was projected to be 8–68 days shorter in 2050, with the most extreme response being recorded for the high carbon dioxide scenario which was simulated with the SILMU (the Finnish Research Programme on Climate Change) global circulation model scenarios (Elo et al., 1998). A later study in Sweden (Lake Erken), using one of the first RCMs (RCA1, developed within the Swedish regional climate modelling programme, SWECLIM) and the A2 emission scenario resulted in a 40 day advance in the mean break-up date of ice in the 2071–2100 period when compared to the historical reference period (1961–1990) with the lake being totally free of ice on two occasions (Blenckner et al., 2002). A more recent study focusing on the same lake, projected a 57–53 days reduction in the median ice break-up date for the 2070–2100 period with the lake beingfree of ice for 2–4 of these years, whereby 4–2 years during the 30 years of our historical study are totally ice free (Persson et al., 2005). Detailed analyses of the temperatures projected for the Northern Lakes during the ice-free period suggest that there will be a significant extension of the period of summer stratification. Further, their results suggested that, depending on the lake, the stratified period in summer will be 23–63 days longer by 2050 than present (1961–1990). The annual surface water temperature was projected to increase 1.8°C (0.5–2.6°C) degrees by 2050 in Lake Pääjärvi. In Sweden, the future climate simulations showed a general increase in water temperature in summer in Lake Erken, with maximum values up to 4°C, compared to the control period, which may lead to a prolongation and a stronger summer stratification (Blenckner et al., 2002; Persson et al., 2005).

In future, the combined effects of the disappearance of the ice and increased water temperatures may mean that some lakes that are currently dimictic lakes (i.e. are fully mixed twice during the year) will become monomictic (i.e. are fully mixed only once during the year from the autumn until the onset of the stratification in the next year) (Schindler, 1996; Blenckner et al., 2002).

18.3 Hydrology

18.3.1 Historical Changes

The variations in the hydrology that associated with the interannual fluctuation in precipitation have important effects on both the water levels and the dynamics of the lakes. Long-term changes in water levels with a periodicity of 20–30 years have, for example, been reported in Peipsi and Võrtsjärv in Estonia (Nõges et al., 2007) and the Müggelsee in northern Germany (Behrendt and Stellmacher, 1987). It is evident that, whilst the primary factor influencing these water levels is the winter rainfall, the strength of the relationship depends on the air temperature, which determines whether the precipitation falls as rain or snow and the subsequent effect on the flow pathways and the absorptive capacity of the soil (Nõges and Nõges, unpublished data). A good example of the effects of systematic changes in the winter precipitation on the dynamics of a lake is that recorded by Nõges et al. (2003) in Lake Võrtsjärv. Lake Võrtsjärv is a large (270 km^2) and shallow (mean depth 2.8 m) lake located in Central Estonia. The lake experiences large Fluctuations in the water level (mean annual amplitude 1.4 m) and the most extreme increases in depth (3.2 m) can increase the surface of the lake by a factor of 1.4 (Nõges et al., 2003).

The fluctuation in the level of Lake Võrtsjärv is strongly related to the total amount of the winter precipitation. In mild winters with increased precipitation, the water level is much higher and remains high until the autumn. In those years, the light conditions as well as phosphorus availability deteriorate, favouring the development of shade-tolerant Filamentous cyanobacteria species like *Limnothrix spp*.

In colder winters with lower water levels, the light penetrates through the whole water column while the phosphorus release increases because of more frequent

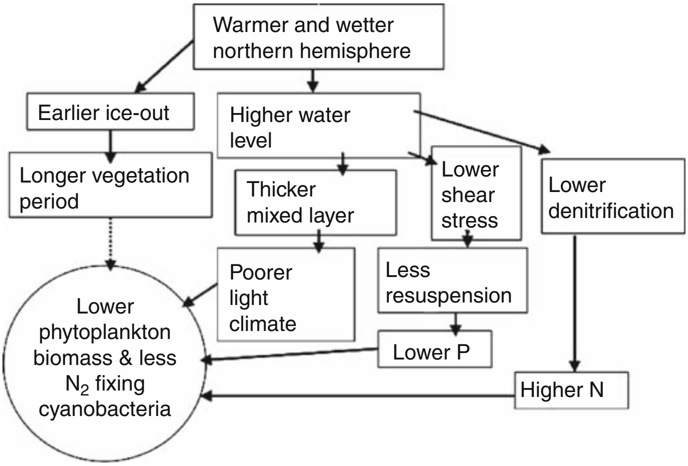


Fig. 18.4 The influence of climate induced variations in the water level on the dynamics of Lake Võrtsjärv in Estonia (Modified from Nõges et al., 2005)

resuspension of bottom sediments, resulting in lower nitrogen to phosphorus ratio. This leads to a lower total phytoplankton biomass but favours nitrogen-fixing cyanobacteria species (Nõges et al., 2003). Therefore, fluctuations in the water level can have a strong influence on the ecosystem dynamics (Fig. 18.4).

18.3.2 Future Projections

The RCM outputs described in Chapter 2, this volume, suggest that there will be a significant seasonal re-distribution of precipitation throughout Northern Europe with the most pronounced changes suggested for the colder half of the year. Decreased snow and ice cover and more frequent periods of partial melting will result in increased winter runoff and earlier snowmelt with lower runoff maxima during the spring. The hydrological projections produced for the CLIME sites (Chapters 9, 11 and 13) highlight this basic pattern. For example, in the Mustajoki sub-catchment of Lake Pääjärvi in Finland, the annual variations in the simulated streamflow were closely correlated with the projected precipitation, i.e. rainfall. Four out of the six scenarios projected increases in annual streamflow for the period of 2071–2100 compared to the control period (1961–1990). There was also an obvious shift in the seasonal pattern of streamflow. In all scenarios the streamflow increased between December and March and decreased between April and October. The main difference in these future simulations is the change in the winter streamflow, i.e. mild winters in the future leads to a more continuous streamflow during the whole winter period due to changes in the snowmelt. In addition, the typical spring maximum in the future flow simulations is 5–7 times lower compared to the control period and was again related to changes in the snowmelt.

18.4 Chemical Responses

18.4.1 Historical Changes

The most important effects of the hydrological changes already described are those associated with the supply of nutrients. In warm and dry years, the flux of nutrients from the surrounding catchment typically decreases but in-lake concentrations may still increase due the increased rate of evaporation, enhance recycling from the sediments or entrainment from deep water. In this section, we first consider the impact of the changing hydrology on the annual loading before describing some more specific seasonal effects.

Long-term changes in the annual loadings of nutrients and dissolved organic matter are often due to the combined effects of local changes in agricultural practices and regional changes in the climate. In CLIME, tests at a number of sites (see Chapters 8 and 10) showed that relatively simple de-trending methods could be used to quantify the short-term (inter-annual) effects associated with changes in the weather. In the Northern Region, the analysis of these de-trended time-series revealed significant positive correlations between the annual loadings of ammonium, phosphate and dissolved organic carbon (DOC) and the long-term change in the measured precipitation (Nõges et al., 2007). The most pronounced internal loading effects are likely to occur in eutrophic lakes that are also thermally stratified. Here, additional phosphorus is released from the sediment during extended period of stable stratification due to the associated variations in the oxygen content, redox potential and microbial activity in the hypolimnion. In contrast, in well mixed, shallow lakes with a short residence times, a high proportion of the phosphorus released from the sediment is lost through the outflow and the effects on the productivity of the lakes are less pronounced. In lakes that stratify intermittently the response can be quite complex since summer warming may increase both the frequency and duration of the individual periods of stratification (Wilhelm et al., 2006).

In seasonal terms, the most important hydrological effects are those reported during the winter/spring period. The analyses of data collated from a number of Swedish and Finnish lakes (Weyhenmeyer, 2004; Järvinen, unpublished data) showed that the winter weather, as quantified by NAO index, had a particularly pronounced effect on discharge, the associated nutrient load and lake chemistry. Mild winters (positive NAOw) can increase the discharge and the associated load of phosphorus and nitrogen in March (Weyhenmeyer, 2004). In these lakes, water chemistry (pH and reactive silica) in May could also be related to changes in water temperature and NAOw (Weyhenmeyer, 2004).

A study of the silicon (Si) cycle in three river-lake systems in Sweden, Estonia and Northern Germany (Nõges et al., 2008) showed coherent seasonality in the Si loads of the two rivers in Sweden and Estonia but the pattern in the German river was rather different. This was probably caused by the common snow-driven hydrology of the catchments in Sweden and Estonia as distinct from the rain-driven hydrology of the German catchment. In all these rivers, the Si load showed a strong seasonal hysteresis where the Si concentrations measured during the rising phase of the hydrograph were higher than those measured during the descending phase.

In contrast to the situation with other nutrients, the coherence of Si dynamics in the lakes depended more on lake morphometry than on the climatic regime: the Si dynamics in the two shallow lakes in Estonia and Germany were more similar than the deeper Swedish lake, even though the Estonian and Swedish lakes had a similar hydrology. Among the variables measured at the three sites, river water discharge was most coherently synchronized by the North Atlantic Oscillation winter index (NAOw). There were significant season-specific correlations of the NAOw with both the biomass and the relative abundance of diatoms in the lakes but there was no coherent pattern among the lakes. The results suggest that processes driven by water discharge are more coherent across regions than in-lake processes.

In the Northern lakes, some of the most important effects of the historical changes in the hydrology were those connected with the leaching of DOC (see Chapter 12). Changes in the temperature, water table and discharge can all affect delivery of DOC to downstream ecosystems, where it can have a direct effect on the attenuation of visible and UV radiation and an indirect effect on the productivity of the lakes and their biogeochemical cycles. Most boreal lakes are CO_2-supersaturated and therefore constitute net sources of CO_2 to the atmosphere. Jonsson et al. (2003) showed that 30–80% of the total organic carbon entering the lakes was lost, mainly due to mineralization and subsequent diffusion of CO_2 to the atmosphere, the exact proportion depending on water residence time and water temperature (Sobek et al., 2003). Since there are at least 80,000 lakes in the boreal zone of Sweden alone, the release of CO_2 from these lakes could, in future, have a considerable effect on the CO_2 concentration of the atmosphere (Kortelainen et al., 2006).

18.4.2 Future Projections

In future, the warmer climate will affect a range of watershed processes: such as reduced soil frost, less snow and snow cover and an earlier start to the growing season. Similarly, increasing amounts of precipitation in winter will increase nutrient leaching from the soil and could accelerate the eutrophication of water bodies. Although the amount of artificial fertilizer used in most agricultural catchments has decreased in recent years, nutrient loads may remain high if there is a sustained increase in the winter run-off (Nõges et al., 2004). The extent of the time-lag between any changes in the catchment and the response of the lakes is difficult to predict. Some investigators (e.g. Karlsson et al., 2005) have suggested that it might take 100–200 years before the vegetation types and soil processes reach a new equilibrium with the climate.

In many lakes, the projected extension of summer stratification is likely to increase hypolimnetic anoxia, which is a precondition for iron-bound and aluminium-bound phosphorus release from lake sediments (Pettersson, 1998). The rates of mineralization in the water column and release from the bottom sediments may also be enhanced by the increased temperatures and the enhanced flux of nutrients to the growing phytoplankton (Blenckner et al., 2002). A recent model study (Malmaeus et al., 2005) used a water quality model driven by a regional climate model to simulate the dynamics of phosphorus in three Swedish lakes. In the two lakes with a short residence time, the projections showed no change in the phosphorus concentration, regardless of lake depth. In contrast, in the lake with a long residence time the annual mean concentration of total phosphorus doubled under the warmest climate scenario. These model projections are discussed in more detail in Chapter 15 and imply that the water quality in eutrophic lakes with long water residence times could deteriorate in the future and suggest that remedial action needs to be taken now to minimise this risk.

The projected changes in the hydrology could also have a significant effect on the supply of allochthonous dissolved organic matter to lakes. Experiments with large mesocosms (Pastor et al., 2003) showed that the dissolved organic carbon (DOC) budget of boreal peatlands was primarily controlled by seasonal changes in the discharge rather than by any effects associated with warming or the observed variations in water level. In a modelling study of the Galten basin in Lake Mälaren, Sweden, (see Chapter 13, this volume) the projected increase in run-off led to a substantial increase in the winter load of DOC, but the simulated increase in temperature had very little effect on the decomposition rates of organic matter.

18.5 Biological Responses

18.5.1 Historical Changes

The biological responses of lakes to the changing climate are often the most complex. Temperature affects all physiological processes and, therefore, has a major effect on the growth of organisms and their survival. Some of the most important effects of the winter climate are those connected with the timing of life history events (Blenckner and Hillebrand, 2002) since phenology is often strongly influenced by temperature and precipitation (Hughes, 2000). In one of the first studies on the impact of climatic events on the plankton, Adrian et al. (1995, 1999) showed that composition, timing and maximum abundance of the phytoplankton and zooplankton populations that develop in the spring were strongly influence by the duration of winter ice-cover. In two German lakes, Heiligensee and Müggelsee, the timing of the phytoplankton minima and zooplankton maxima (the so called clear-water phase) was negatively correlated with the water temperature, i.e. increases in the temperature led to an earlier clear-water phase. In each case, the size of the spring algal maximum was negatively correlated with its timing. Similarly, the timing of the spring phytoplankton bloom in Lake Erken in Sweden was inversely related to the temperatures experienced during the winter (Weyhenmeyer et al., 1999), i.e. earlier spring blooms of phytoplankton were associated with mild winters and an earlier break-up of lake ice. This study (Fig. 18.5) covered an exceptionally long

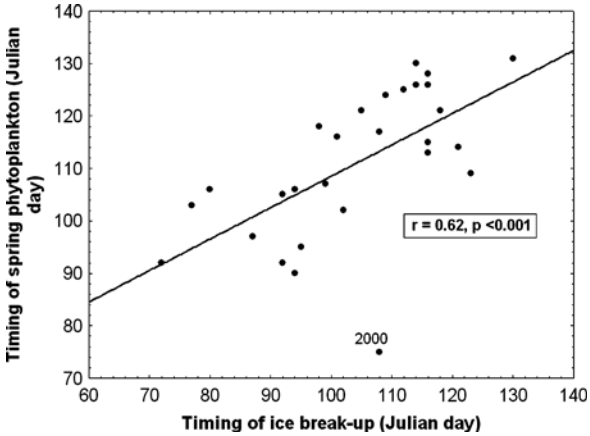


Fig. 18.5 The relationship between the timing of the spring phytoplankton bloom in Lake Erken and the timing of ice break-up Julian day, 1. Jan = 1

period of time (1941 and 2005). Although there were some gaps in the record, the only obvious outlier was 2000. In that year, a substantial growth of the phytoplankton was recorded under the ice, due to the enhanced penetration of light brought about by the lack of snow.

The results from Swedish lakes (Weyhenmeyer, 2001) and Lake Pääjärvi, Finland (Järvinen et al., 2006) suggest further that the temperature-sensitive phytoplankton groups, cyanobacteria and chlorophytes, would benefit from the earlier warming of the lakes and the earlier onset of thermal stratification. At all these sites, there was a significant correlation between their biomass in May and June and the recorded inter-annual variation in the spring temperatures. In addition, water temperatures in summer above ~20°C could clearly favour cyanobacteria blooms (Nõges et al., unpublished).

The growth and survival of most fish species is strongly dependent on temperature (Magnuson et al., 1990; De Stasio et al., 1996; Magnuson et al., 1997). Temperature-induced changes in the growth rate of predatory fish can also give rise to cascading effects through the entire food web (Carpenter et al., 1985). In shallow lakes, the effects of variations in the winter climate on plankton are usually short-lived and are soon overtaken by the biotic interactions that regulate these populations in the spring. More persistent effects may, however, appear in lakes where there are winter fish-kills and where significant changes in the structure of the food web can still be detected after several years (Adrian et al., 1999). In deep lakes, however, the effects of winter warming, such as those associated with the NAO, may persist until late summer (Gerten and Adrian, 2001).

18.5.2 Future Projections

Future projections of biological changes in lakes present an enormous challenge to our current understanding of many key processes. Two complementary approaches have been adopted to explore these effects i.e. experiments and modelling. A good example of the experimental approach is that described by Kankaala et al. (2000) and Ojala et al. (2002). These investigators designed a 3-year study of the impact of increased temperatures and elevated $C0_2$ on the growth of phytoplankton and emergent macrophytes in two experimental ponds in southern Finland. Both experiments were carried out in a plastic greenhouse. The results showed that the emergent macrophytes grew earlier and better in the warmer conditions (+2.5–3°C). The enhanced released of phosphorus from decaying plants in the greenhouse pond also stimulated the growth of more filamentous algae. Such results demonstrate the key role played by macrophytes in the uptake of phosphorus and control of eutrophication in these small lakes and suggest that the future management of lakes needs to pay particular attention to the fate of macrophyte communities.

An experiment with natural plankton communities grown at 10 and 20°C (Rae and Vincent, 1998) showed that increasing temperatures could also have a major effect on the structure of the plankton community. Warming is likely to accelerate eutrophication processes in northern lakes, thereby shifting the composition of the phytoplankton toward bloom-forming groups, like the cyanobacteria, that have higher temperature optima.

Under the warmer conditions projected for the future, cyanobacteria may become increasingly dominant in many lakes since the growth of buoyant, bloom-forming species is favoured by warm and stable water conditions (Visser et al., 1996). A good example of the holistic response of a lake to a change in the regional climate is that described in Chapter 15 for Lake Erken in Sweden. Here, a biogeochemical model (BIOLA) was used in conjunction with a transient simulation of the regional climate to quantify the year-on-year changes projected between 1961 and 2100. In these simulations, the annual water temperature increased by around 1°C, the summer period of stratification was both longer and more intense, there was a marked decrease in the hypolimnetic oxygen concentration and a general increase in the biomass of phytoplankton. Very similar trends were recorded in a large polymictic lake (the Galten basin of Lake Mälaren) using a phytoplankton model linked to a hydrodynamic model and a nutrient flux model for the surrounding catchment (Markensten, 2005). In their future climate mode, these models projected an increase in the number of stratification events, an increase in the biomass of phytoplankton and a shift away from diatoms towards a dominance of cyanobacteria.

18.6 Impacts of Non-climatic Anthropogenic Drivers

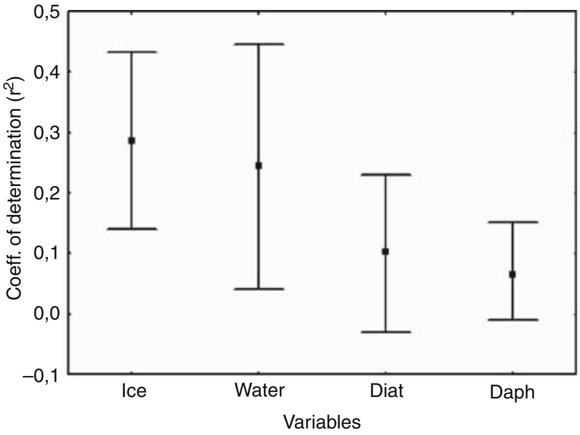
In the past 40–50 years, many lakes, in Europe and North America, have been subjected to an anthropogenic increase in the supply of nutrients and other substances (e.g. organic pollutants). This supply has decreased somewhat since the 1980s due to improvements in wastewater treatment but the supply of nutrients from diffuse sources is still a source of concern. Other human influences such as land-use changes, the construction of dams and the effects of acidification further, complicate the separation of 'natural' and anthropogenic influences on ecosystems (Andersson and Arheimer, 2003; Jeppesen et al., 2003; Van Donk et al., 2003; Köhler et al., 2005). It is therefore often difficult to distinguish between the regional effects attributable to the climate and the more local effects of direct human impacts. Some systems respond in quite a complex way to the combined effects of a number of external drivers. In one recent study, from Müggelsee in Northern Germany, simulations showed that reductions in the supply of phosphorus can counter the effects of reduced ice-cover by delaying the spring diatom bloom and introducing a 'switch' from a bottom-up to a top down mode of regulation (Huber et al., 2008). Multiple 'pressure' factors, such as climate change, ozone depletion and acidification can also alter ultraviolet (UV) light and temperature regimes in freshwater ecosystems. Phytoplankton communities experience climate change indirectly through changes in lake level, timing of ice break-up, stratification, nutrient inputs and zooplankton grazing (Anneville et al., 2004).

18.7 Discussion

In Northern Europe, as elsewhere, the most pronounced limnological effects were observed with the physical variables, like water temperature and the mixing regime, that were directly driven by the observed and projected changes in the climate. In winter the strongest and most coherent responses to the changing climate were those connected with the reductions in the extent and duration of ice cover. A comprehensive analysis of these winter effects has been produced by Blenckner et al. (2007) in a meta-analysis of the physical, chemical and biological data acquired from eighteen European lakes. Here, the strongest overall response to the winter climate (as measured by the NAOw) was exhibited by the water temperatures but significant effects were also noted for the summer biomass of cyanobacteria, the spring abundance of daphnids and the summer abundance of the larger zooplankton. Figure 18.6 shows the mean, minimum and maximum of the coefficients of determination for four of the variables analysed in this study. The statistics are based on the calculated correlations between the selected variable and the NAOw. All the lakes were located in Northern Europe and included Lake Pääjärvi in Finland, Lake Erken, and the Galten and Ekoln basins of Lake Mälaren in Sweden and Lake Võrtsjärv in Estonia. These results highlight the fact that the strongest responses are found in the timing of ice break-up (mean $r^2 = 0.29$) and surface water temperature in May (mean $r^2 = 0.25$). The response of the diatom biomass in May and the abundance of daphnia in June are not so clear since they are more likely to have been influenced by lake specific factors such as the supply of nutrients, water colour, biological interactions.

In a comparable study of the deep water temperature of twelve European lakes that included two Swedish lakes, Dokulil et al. (2006) showed that the pattern of warming was highly coherent (see also Chapters 4 and 17 of this volume). The average rate of hypolimnetic warming was 0.1–0.2°C per decade and the rate for the two Swedish lakes was 0.15°C per decade.

Fig. 18.6 The median, min and max coefficient of determination of the relationship between the NAOw and the timing of ice break-up (Ice), May surface water temperature (Water), the May diatom biomass (Diat) and the early June Daphnia biomass (Daph) in the Northern European lakes



Future climate change will further affect the stratification regime of lakes and many lakes that are currently dimictic lakes could become monomictic or temporarily stratified. Also some meromictic and polymictic lakes may become dimictic (or even monomictic if they are totally ice-free). In general, high summer water temperatures will lead to longer periods of summer stratification and increased water column stability which, in turn, could give rise to increased hypolimnetic anoxia, or at least greatly reduced oxygen concentrations (Magnuson et al., 1997). Such changes could have far reaching ecological consequences if they substantially increase the rate at which nutrients are release from the surficial sediment (Pettersson et al., 2003; Wilhelm and Adrian, 2006).

The schematic in Fig. 18.7 summarizes the key climate-related factors that have already influenced the dynamics of lakes in Northern Europe and includes an assessment of the confidence placed in the specified interactions. In future, the most important effects are likely to be those associated with the change in temperature, i.e. the growth rate of the organisms, the recycling of nutrients and lake productivity. Some of these responses could well be quite complex, particularly those connected with the phenology of life history, their stochastic interplay and the non-linear ‘threshold’ that often characterize these lake ecosystems.

However, the magnitude of change will also be strongly influenced by lake specific features, i.e. its geographic position, the nature of the surrounding catchment, the morphometry of the lake, various anthropogenic influences (e.g. eutrophication) and abiotic and biotic interactions (Blenckner, 2005), which all contribute to poor predictability. This underlines the fact that management actions designed to mitigate climatic impacts, like increases in the water colour or internal eutrophication, need to be based on the characteristics of the individual lakes and their surrounding catchments. In all cases, the challenge will be to develop more adaptive methods of management that promote high tolerance to multiple pressures in the selected ecosystem, i.e. increase the resilience of the ecosystem (Janssen and Carpenter, 1999; Folke et al., 2004). Model predictions that reach a better certainty or, in other words, a

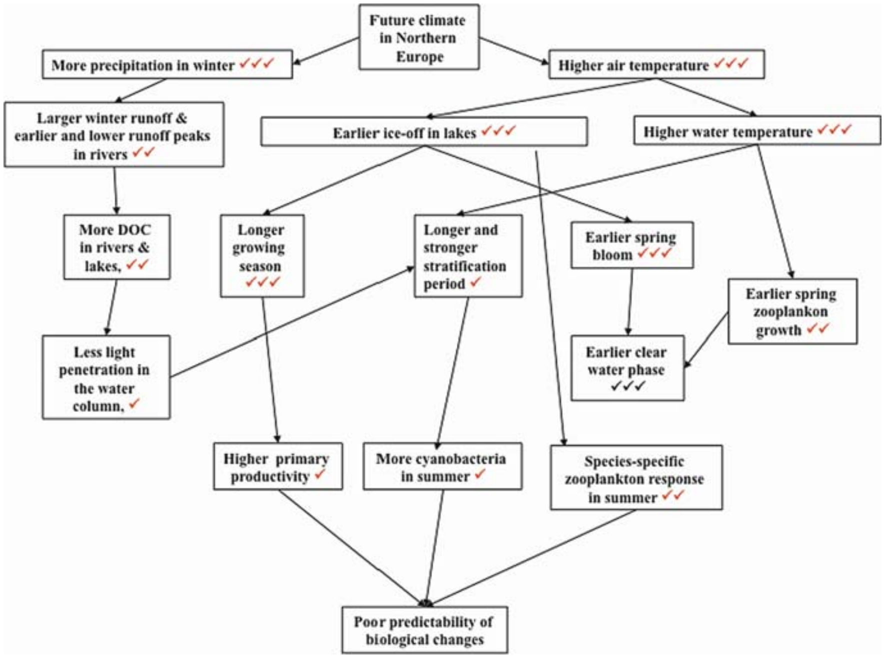


Fig. 18.7 Schematic overview of the main climatic responses expected in Northern European lakes. The number of ticks (√) indicate the level of confidence in a particulate response

lower risk of failure are feasible, a management towards a high stress tolerance remains necessary. Therefore, ecosystems should be managed in an adaptive and precautious way based on the evaluation of the ongoing monitoring programme in order to maintain natural resilience and thus be prepared for future changes.

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Chapter 19
The Impact of Climate Change on Lakes in Britain and Ireland

Glen George, Eleanor Jennings, and Norman Allott

19.1 Introduction

The CLIME studies in Western Europe were confined to a relatively small number of lakes situated in Britain and Ireland. In this chapter, we use the terms 'Britain' and 'Ireland' when describing the location of the lakes, and the term 'British Isles' when referring to the area covered by the weather typing system devised by Lamb (1950). The climate of Britain and Ireland is notoriously variable and is strongly influenced by the movement of weather systems across the Atlantic (Barrow and Hulme, 1997). Winters are typically mild and wet but there are large year-to-year variations in the seasonal distribution of the rainfall. The variable nature of the climate has a pronounced effect on the physical, chemical and the biological characteristics of the lakes (Allott, 1986; George et al., 2004). In this chapter, we review the climatic changes projected for the region and assess their potential impact on the characteristics of the lakes. The review is based on the climate change projections described in Chapter 2 and an analysis of the variations recorded in a number of intensively studied sites. Most of the lakes covered by this review are relatively deep and are thermally stratified throughout the summer. Their climatic responses are, consequently, very different from those observed in the polymictic lakes that form part of the Acid Waters Monitoring Network (www.ukawmn.ucl.uk) or the shallow lakes of the Norfolk Broads (Moss et al., 1996). To simplify the review, we confine our attention to the effects reported during the winter (November, December and January) and summer (June, July, August). In winter, weather patterns in Britain and Ireland are strongly influenced by the atmospheric feature known as the North Atlantic Oscillation (Hurrell et al., 2003). In summer some of the most important effects are those associated with the north-south movements of the Gulf Stream (Taylor, 1996) and the changing frequency of warm, anticyclonic days (Kelly et al., 1997).

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Several accounts of the impact of climate change on water supplies in the region have already been published (Arnell, 1999; Arnell and Delaney, 2006; Charlton et al., 2006). Less attention has hitherto been paid to the effect of these changes on the quality of the water and the seasonal dynamics of different types of lakes. Notable exceptions include the long-term studies on the lakes of the English Lake District (George et al., 2000; George et al., 2004; George et al., 2007b), observations on two lakes in the west of Ireland (Jennings et al., 2000; Allott and Jennings, 2006) and the microcosm experiments described by McKee et al. (2003). In this chapter, we highlight those aspects of CLIME that are most relevant to lake and catchment managers in Britain and Ireland. More detailed accounts of the patterns and processes involved can be found elsewhere in the volume, notably in Chapters 10, 14, 15 and 16.

19.2 The Distribution of Lakes in Britain and Ireland

The map in Fig. 19.1 shows the location of the main lake districts in Britain and Ireland. In Britain, the highest concentration of lakes is found in the North of Scotland, the North West of England and North Wales. Scotland alone has more than 8,000 lakes (lochs) that are larger than one hectare in area (Hughes et al., 2004; Maitland et al., 1994). Accounts of the physical, chemical and biological

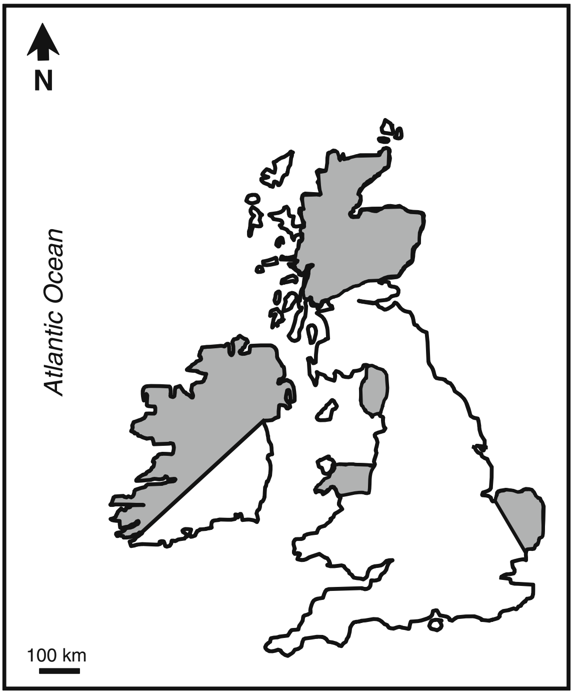

Fig. 19.1 The location of the main lake districts in Britain and Ireland

characteristics of the larger lochs in Scotland have been given by Maitland (1981) and George and Jones (1987). The most intensively studied loch in Scotland is Loch Leven, a shallow loch that is known to be very sensitive to changes in the rainfall (Bailey-Watts et al., 1990; Bailey-Watts and Kirika, 1999). In England, the sites that have been the subject of the most intensive studies are located in the English Lake District (Macan, 1970; Talling, 1999). Most of these lakes are deep and their trophic status is largely determined by the geology and soils of the surrounding catchments. The shallow lakes that form the Norfolk Broads and the Cheshire Meres have also been a focus of scientific investigation. A review of the ecology of the Broads was published by Moss (2001) and an account of a recent study of Rostherne Mere by Krivtsov et al. (2001). The largest lakes in Wales are those located in Snowdonia (Eryri) but there are also a number of reservoirs in the central uplands. Most of the lakes in Ireland tend to be shallow for their surface area (Reynolds, 1998). The deepest lakes are located in the mountains to the south and west but there are some relatively deep lakes on the River Shannon in the midlands. Many lakes in Britain and Ireland are used as sources of drinking water and most of the lake districts are now important centres for tourism and recreation.

19.3 The Climate of Britain and Ireland

The climate of Britain and Ireland is strongly influenced by the warm waters of the North Atlantic Drift. The mildest areas are located along the south and west coasts and the coldest areas in the Scottish mountains. Mean annual air temperatures range from a minimum of 5°C in the north to a maximum of 12°C in the south west. Cold winters, with average temperature below 2°C are rare, but warm summers, with average temperatures above 15°C are relatively common. Precipitation gradients throughout the region are very steep and heavily influenced by the local topography. In the western mountains, the annual rainfall can exceed 3,000 mm (Allott et al., 2008). In the southern and eastern lowlands, the annual average is typically less than 1,000 mm per year.

The key factor regulating the winter weather is the atmospheric feature known as the North Atlantic Oscillation (NAO). Positive values of the index used to quantify this pressure gradient (the NAOI) are associated with mild, wet winters and negative values with colder, drier conditions (Jennings et al., 2000; George et al., 2004). Air temperatures throughout the region are positively correlated with the NAOI but the rainfall correlations are more variable with strong positive correlations only being observed in western areas. Figure 19.2 shows the effect that year-to-year variations in the NAOI had on the geographic variation in the rainfall. In the north and the west, there was a significant positive correlation between these variations and the NAOI. In the east of Ireland, the correlations were not significant whilst negative correlations were commonly reported from sites in the east of England (Wilby et al., 1997).

Summer weather patterns in the region are not greatly influenced by the NAO but are affected by the trajectory of storms in the Atlantic (Taylor, 1996) and with

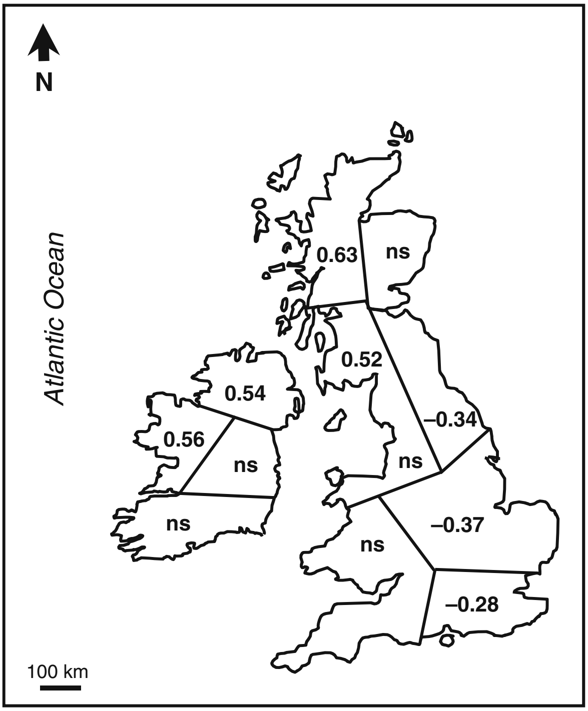

Fig. 19.2 The correlations between the average winter rainfall and the NAOI for different areas of Britain and Ireland (modified from Wilby et al., 1997). ns = not significant

the number of anticyclonic days when the area is dominated by high pressure (Kelly et al., 1997). Some examples of these high pressure effects are given in George et al. (2007a) and in Chapters 8, 10 and 16 of this volume. In Britain and Ireland, these periods of high pressure have a profound effect on the summer dynamics of the lakes. For example, the lower wind speeds associated with these anticyclonic situations not only favours the growth of cyanobacteria but can also give rise to 'bloom' conditions when the algae accumulate downwind (George, 1992). These changes in the circulation of the atmosphere are now known to be closely correlated with the north–south movements of the Gulf Stream in the western Atlantic. In winter, the influence of the Gulf Stream greatly reduces the severity of the weather experienced in the south west of Ireland (Allott, 1986). In summer, variations in the position of the Gulf Stream regulate both the mixing characteristics of lakes in the English Lake District (George and Taylor, 1995) and the supply of nitrate to two lakes in the south west of Ireland (Jennings and Allott, 2006).

19.4 The Climate Change Scenarios for Britain and Ireland

In situations where it is difficult to predict the future, it is customary to resort to scenario based analyses. A climate change scenario is an internally-consistent description of how the climate might change given a specified increase in the greenhouse gas concentration. In CLIME, we used the downscaled outputs from two Global Climate Models (the Hadley Centre model and the Max Planck ECHAM model), each driven by two contrasting emission scenarios (IPCC: A2 and B2) to inform our assessments. Details of these models are given in Chapter 2, which also discusses the uncertainties associated with the projections produced for the different regions. In this chapter, we confine our attention to the changes projected for winter and summer, the two seasons that have the most significant effects on the physical, chemical and biological characteristics of the lakes.

Figure 19.3a and b show the change in the winter rainfall projected by the two models for Britain and Ireland. The Hadley Centre model (Fig. 19.3a) suggests that the rainfall will increase by an average of 20% with the most pronounced increases occurring in southern Britain. The increases suggested by the Max Planck model (Fig. 19.3b) are much larger and also cover a larger area. Figure 19.3c and d show the change in the summer air temperatures projected by the two models. The Hadley Centre model (Fig. 19.3c) suggests that air temperatures will increase by an average of 3°C. The increases suggested by the Max Planck model (Fig. 19.3d) are larger and also cover a larger area. Related studies, not described here, suggest

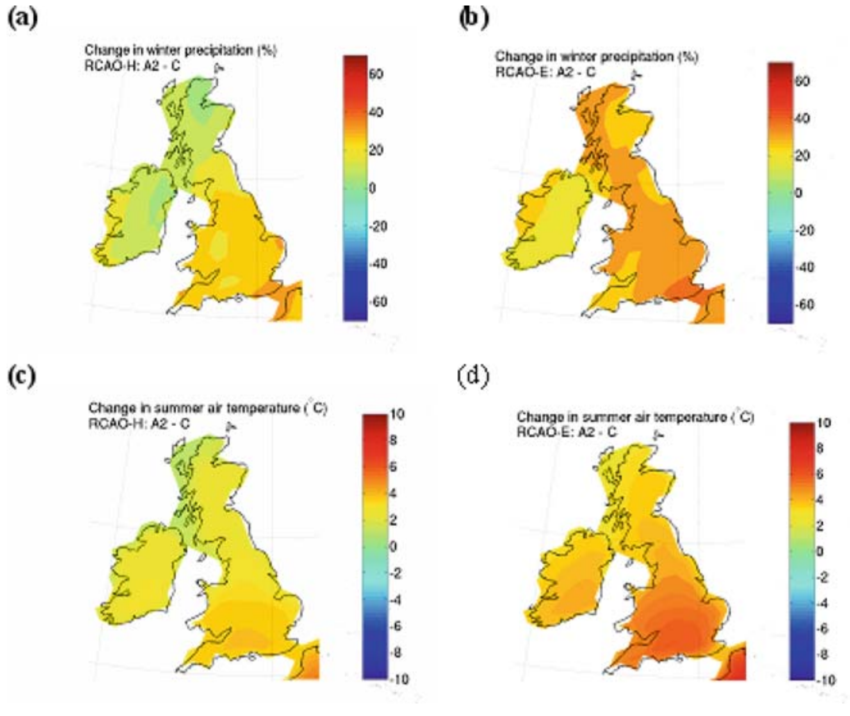

Fig. 19.3 The change in the winter rainfall projected by (**a**) the Hadley Centre model (RCAO-H) and (**b**) the Max Planck model (RCAO-E). The change in the summer air temperature projected by (**c**) the Hadley Centre model and (**d**) the Max Planck model. All projections are based on the IPCC A2 scenario for the period between 2070 and 2100

that the daily weather experienced in the area will also become more variable in a warmer world. For example, Hulme et al. (2002) suggest that periods of very heavy rain will become more common in winter and that there will be a marked increase in the number of very hot summer days. In the impact assessments that follow, we used these projections together with the results summarized in Chapter 2 to make the following assumptions about the future climate of Britain and Ireland:

- Winters will become milder and wetter with the most pronounced changes recorded in the north and west.
- Summers will be warmer and drier with the most pronounced changes recorded in the south and east.
- Extreme weather events, such as heavy winter rains and prolonged summer droughts, will become increasingly common.

The changes projected for other climatic variables are less certain. The UKCIP scenarios (Hulme et al., 2002) suggest that there will be a general decrease in cloud cover during the summer and a corresponding reduction in the relative humidity. The most pronounced changes are those projected for southern Britain where the relative humidity could be at least 10% lower than it is today. The number of days when the islands experience prolonged periods of high pressure in summer is also likely to increase and would lead to a marked reduction in the wind speeds recorded throughout the area.

19.5 The Impact of Mild, Wet Winters

19.5.1 The Physical Effects of Mild, Wet Winters

Ice is rare on Irish lakes and most lakes in Britain only freeze for a few days in the year. As winters become milder, lowland lakes are likely to remain free of ice but lakes situated at higher altitudes will still freeze during the coldest winters. Long-term ice records are rare in Western European lakes but George (2007) has shown that there has been a progressive decline in the number of 'ice-days' recorded on Windermere. In the 1970s, the number of days when some ice was recorded in a sheltered bay averaged 10.1 days per year but this declined to 1.7 days per year in the 1990s. In lakes that are free of ice for most of the winter, the surface temperature is closely correlated with the local air temperature. The relationship between the two variables is, however, influenced by a number of site-specific factors, such as the depth of the lake and its exposure to the wind. Model simulations in the English Lake District (George et al., 2007b) demonstrate that the largest increases in the surface temperature (+1.1°C) will occur in the shallowest lakes and the smallest increases (+0.5°C) in the deep lakes with very long residence times.

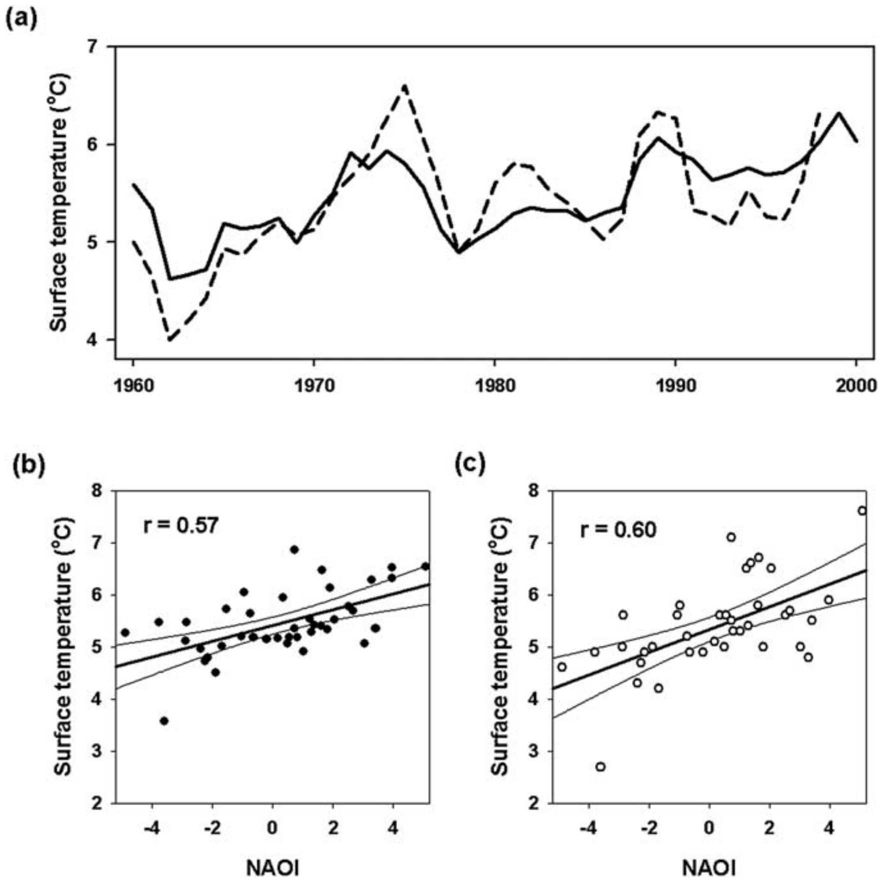

Fig. 19.4 (**a**) The year-to-year variations in the winter surface temperature of the North Basin of Windermere (–) and Lough Feeagh (- -). The time-series have been smoothed with a three-point running mean. (**b**) The relationship between the winter surface temperature of the North Basin of Windermere and the North Atlantic Oscillation Index (NAOI). (**c**) The relationship between the winter surface temperature of Lough Feeagh and the NAOI

The most detailed lake temperature records available for Britain and Ireland are those collated for Windermere in the English Lake District and Lough Feeagh in the west of Ireland. At both sites, the surface temperature has been measured at daily intervals and the observed variations related to the long-term change in the regional weather (George et al., 2007b). The North Basin of Windermere (54°22'N; 2°56'W) covers an area of 8.1 km^2 and has a maximum depth of 60 m. Lough Feeagh (53°50'N; 9°35'W) has a surface area of 3.9 km^2 and a maximum depth of 45 m. The surface temperature measurements for Windermere were taken at ca 08.00 hours GMT using a mercury thermometer with a nominal resolution of 0.1°C. The corresponding values for Lough Feeagh were abstracted from paper charts produced by an electro-mechanical recorder. The year-to-year variations in the winter temperatures of the two lakes (Fig. 19.4a) show that these have increased progressively over the years. The regression lines fitted to the original measurements showed that the trends were significant at the 2% ($p < 0.02$) level. The average rates of increase were 0.22°C per decade for Windermere and 0.25°C per decade for Lough Feeagh. Much of this warming can be related to the inter-annual variations in the NAO. In Windermere (Fig. 19.4b), the correlation between the average winter temperature and the NAOI was 0.57 ($p < 0.001$). In Lough Feeagh (Fig. 19.4c) the correlation was 0.60 ($p < 0.001$). In both lakes, the highest winter temperatures were recorded in the early 1990s when the NAOI remained in its positive 'mild winter phase' for five consecutive years.

The effect of the projected increases in the winter rainfall on the residence time of a lake depends on its size, catchment area and geographic location. The most sensitive sites will be small lakes located in the 'wet' west and the least sensitive sites large lakes situated in the 'dry' east. George and Hurley (2003) used a hind-casting procedure to quantify the seasonal variations in the residence time of the larger lakes in the English Lake District. The results showed that some lakes behaved as 'one season' lakes that responded very quickly to changes in the rainfall. Others could best be described as 'multi-season' lakes, where the effects of any change in the rainfall extended over periods that ranged from several months to more than a year. The most sensitive lake in the Lake District was Rydal Water, a lake with an average winter residence time of 10 days. The least affected was Wastwater, a lake with an average winter residence time of 435 days. Many large lakes in Ireland have relatively short residence times. For example, Lough Leane has an average winter residence time of 124 days whilst the corresponding estimate for Lough Feeagh is 110 days. Ireland also has more than 250 temporary lakes (turloughs) situated in the western limestone region. Turloughs normally become flooded in late September or early October and dry out by the following June. Water levels in turloughs fluctuate dramatically in the winter in response to changes in the local rainfall. The projected increase in the winter rainfall is thus likely to lead to higher maximum water levels in these turloughs with an associated risk of severe flooding.

19.5.2 The Chemical Effects of Mild, Wet Winters

Milder winters will, in general, increase the rate at which minerals and nutrients are leached from rocks and soils in the catchment but a few chemicals, such as aluminium hydroxide, become less soluble as the temperature increases. The impact of the projected changes in the weather on the leaching and transport of nutrients is quite difficult to assess. Much depends on the frequency of heavy rains, the nature of the soil and the distribution of point and diffuse sources within a catchment. Heavy rains dilute point sources of pollutants but may increase the total solute load from diffuse sources. Meyer and Likens (1979) showed that the annual export of phosphorus from a New Hampshire catchment was a linear function of the average stream-flow. In contrast, winter phosphate inputs into Loch Leven (Bailey-Watts et al., 1990) were not correlated with the rainfall since most of the phosphate was derived from point sources. In recent years, there has been a steady increase in the concentration of nutrients reaching most lakes in Britain and Ireland. In Britain, these trends are principally related to the increased loading of sewage treatment works situated in both rural and urban locations. Many of these sewage treatment works are now equipped with tertiary treatment plants but the efficiency of these systems is inversely related to the volume of effluent processed (Pickering, 2001). In Ireland, a country dominated by dairy and beef farming, these trends are more likely to be associated with the increased use of artificial fertilizers and the periodic application of slurry (e.g. Tunney et al., 1998; Zhou et al., 2000; Bunting et al., 2007).

Anthropogenic influences of this kind greatly complicate the analysis of historical data acquired over several decades. In most cases, however, the time-series can be de-trended and the residuals used to quantify the short-term effects associated with the year-to-year variations in the weather. George et al. (2004) used this approach to quantify the effects of changes in the winter weather on the winter concentrations of dissolved reactive phosphorus and nitrate-nitrogen in four English lakes. Two climate-related effects were identified in the de-trended time-series:

- A rainfall-related increase in the de-trended concentration of dissolved reactive phosphorus which was only statistically significant in the small lakes with shorter residence times.
- A temperature-related reduction in the de-trended concentration of nitrate-nitrogen which was statistically significant in all the lakes, irrespective of size.

Figure 19.5a shows the inter-annual variation in the de-trended concentrations of dissolved reactive phosphorus (DRP) and nitrate-nitrogen (NO_3-N) in Blelham Tarn, a small productive lake in the Windermere catchment. The concentrations shown are the averages of the measurements taken in the first ten weeks of each year. These averages provide the best measure of ‘winter’ conditions since the water temperature is then close to the seasonal minimum and there is very little biological uptake. Once the long-term trend has been removed, the residual variation in the DRP was closely correlated with the observed variation in the winter rainfall. The highest de-trended concentrations were always recorded in wet winters and the lowest in winters when there had been very little rain (Fig. 19.5b). The most likely explanation for this variation is the effect that heavy rains have on the routing of surface drainage within the catchment (Sharpley and Syers, 1979). If the rainfall is light, much of the dissolved phosphorus is adsorbed as the water passes through the soil. Heavy rains increase the proportion of water reaching the streams as an overland flow which typically contains much higher concentrations of DRP (McDiffet et al., 1989).

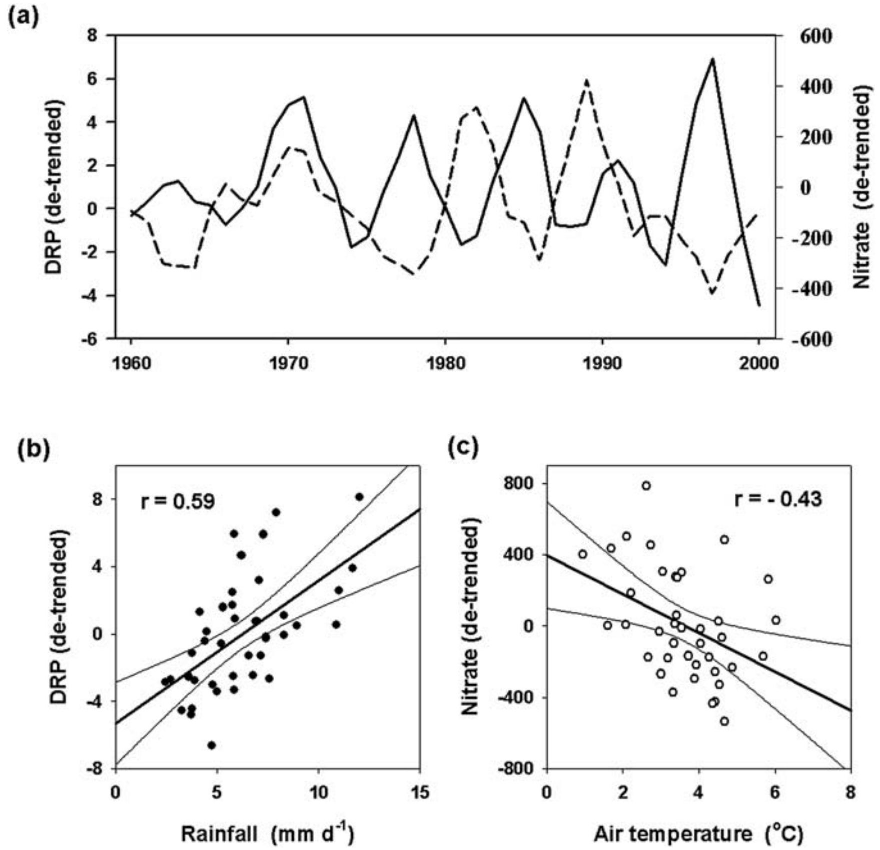

Fig. 19.5 (**a**) The year-to-year variations in the de-trended winter concentration of dissolved reactive phosphorus (DRP —) and nitrate-nitrogen (NO3-N - -) in Blelham Tarn. The time-series have been smoothed with a three-point running mean. (**b**) The relationship between the de-trended DRP and the winter rainfall. (**c**) The relationship between the de-trended nitrate and the winter air temperature

The inter-annual variations in the residual concentration of nitrate-nitrogen (Fig. 19.5c) followed a very different pattern. Here, the highest de-trended concentrations were recorded in cold winters and the lowest when the winters were mild. In contrast to the situation with the DRP, the nitrate-nitrogen fluctuations followed the same pattern in all the lakes irrespective of their size and trophic status. George et al. (2004) suggested that the critical factor was the increased terrestrial assimilation of nitrate in mild winters. Microbial de-nitrification is known to influence the winter concentration of nitrate in wet soils (Groffman and Tiedje, 1991) but some nitrate may also be assimilated by the terrestrial vegetation in very mild winters (Lain et al., 1994). More detailed accounts of the factors influencing the supply and re-cycling of nutrients in a number of different lakes are given in

Chapters 8 and 10 whilst Chapters 9 and 11 describe some of the associated modelling studies.

19.5.3 The Biological Effects of Mild, Wet Winters

In most lakes, the winter growth of phytoplankton is controlled by the availability of light (Talling, 1971) so the increased cloud cover associated with the projected increase in the rainfall could limit the attainable biomass. The direct effects of the projected increases in the air temperature are likely to be less important than their indirect effects on the mixing regime and the underwater light climate. In lakes with short residence times, the flushing effect of the projected increases in the rainfall could, however, have a significant effect on the growth of phytoplankton. Figure 19.6a shows the effect that the year-to-year variations in the rainfall, recorded between 1968 and 2000, had on the winter biomass of phytoplankton in Blelham Tarn. Blelham Tarn has an average winter residence time of 28 days but this can decline to an average of 16 days in very wet winters (George et al., 2007b). In lakes with long residence times these flushing effects are more difficult to detect. Significant shifts have, however, been observed in the timing of the spring bloom, which can be delayed by several days if the winter has been very wet. In 2003, George and Hurley used a Gaussian model to quantify the effects of changes in the weather on the seasonal dynamics of phytoplankton in a number of English lakes. The timing of the spring chlorophyll maximum was measured by the Mμ 1 parameter, an index that ranges from a minimum of 0.17 (Julian day 62) to a maximum of 0.28 (Julian day 105). Figure 19.6b shows the relationship between this measure of the timing of the spring maximum and the corresponding variation in the winter residence time

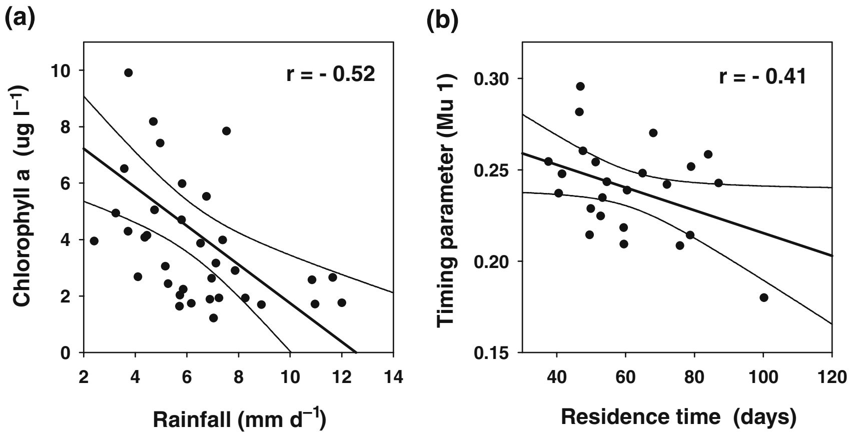

Fig. 19.6 (**a**) The impact of year-to-year variations in the rainfall on the winter biomass of phytoplankton in Blelham Tarn. (**b**) The impact of year-to-year variations in the winter residence time on the timing of the spring biomass maximum in Esthwaite Water

of Esthwaite Water. The results demonstrate that these variations had a significant effect on the timing of the spring bloom ($r = 0.41$, $p < 0.05$) even though this lake has an average winter residence time of 64 days. In the 1970s, the average value of the $M\mu 1$ index for Esthwaite Water was 0.24 (Julian day 87) but this changed to an average of 0.27 (Julian day 99) in the 1990s.

19.6 The Impact of Warm, Dry Summers

19.6.1 The Physical Effects of Warm, Dry Summers

In summer, the most important weather-related effects are those associated with the increase in the air temperature, the reduction in cloud cover and the variation in the wind speed. In thermally stratified lakes, the vertical transfer of heat is controlled by the combined effects of the air temperature, the attenuation of light and the frequency and intensity of wind-induced mixing. The summer surface temperatures of all the lakes analysed in Britain and Ireland have increased progressively over the last 40 years (Chapter 6, this volume). In Windermere and Lough Feeagh (Fig. 19.7a), the long-term variations in the surface temperature have followed a quasi-cyclical pattern with high values being recorded in 1968, 1976, 1983 and 1989. Regression lines fitted to the raw time-series showed that these trends were significant at the 2% ($p < 0.02$) level. The average rate of increase for Windermere was 0.40°C per decade whilst that for Lough Feeagh was only 0.25°C per decade. Much of this increase can be related to a change in the number of anticyclonic days recorded over the British Isles during the summer. In Windermere (Fig. 19.7b) the correlation between the surface temperature and the number of anticyclonic days was $r = 0.60$ ($p < 0.001$). In Lough Feeagh (Fig. 19.7c) the correlation was $r = 0.59$ ($p < 0.001$).

The effects of the associated variation in the summer rainfall are more site-specific and depend on the size and the residence time of each lake. In isothermal lakes, the residence-time is usually a simple function of the total volume. In thermally stratified lakes, the residence time is controlled by the depth of the thermocline and by the onset and duration of thermal stratification. George et al., (2007b) estimated that the summer residence time of some small lakes in the Lake District could increase by as much as 50% in the next 50 years but no estimates are available for any lakes in Ireland.

19.6.2 The Chemical Effects of Warm, Dry Summers

Weather-related effects on the external supply of nutrients are difficult to detect in summer since any nutrients that reach the lakes are rapidly depleted by the growth of phytoplankton. In small lakes, significant inter-annual variations in the residual concentration of nitrate have, however, been detected and related to changes in the intensity of wind-mixing (Chapter 10, this volume). In large lakes, these

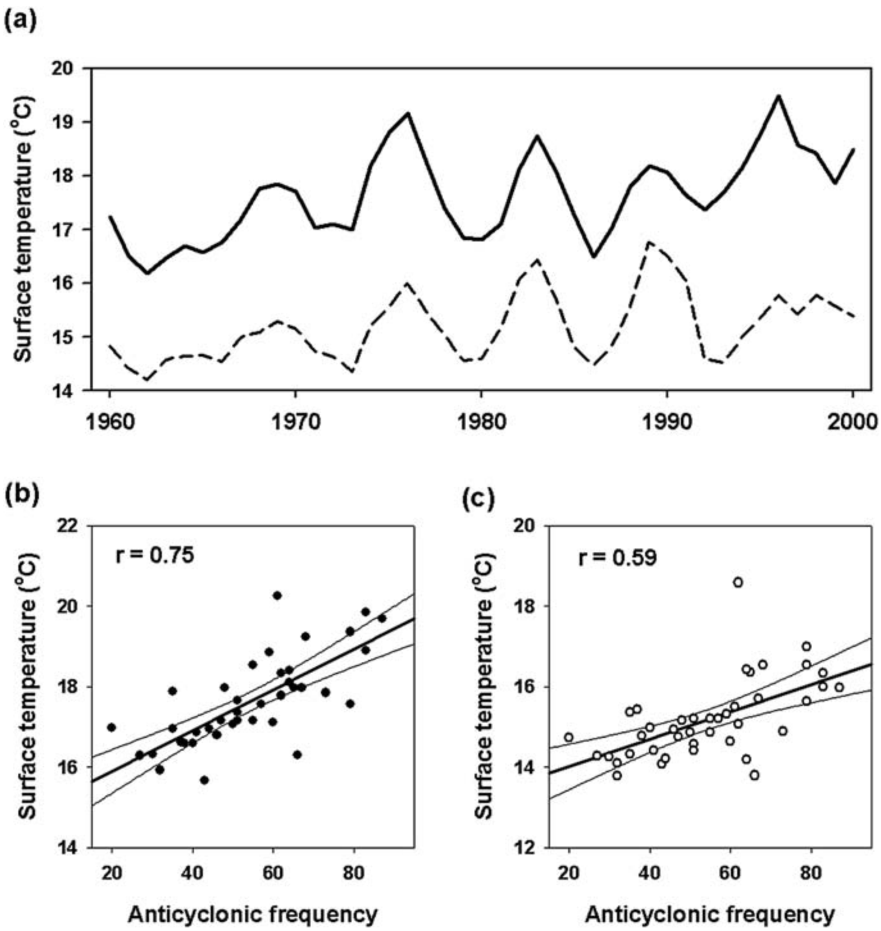

Fig. 19.7 (**a**) The year-to-year variations in the summer surface temperature of Windermere (—) and Lough Feeagh (- -). The time-series have been smoothed with a three-point running mean. (**b**) The relationship between the summer surface temperature of Windermere and the number of anticyclonic days. (**c**) The relationship between the summer surface temperature of Lough Feeagh and the number of anticyclonic days

variations are usually masked by the integrating effects of the catchment and the long residence times of the individual basins. Increases in the supply of nitrate following unusually dry summers have, however, been noted in a number upland catchments (Reynolds et al., 1992) and catchments subject to intensive cultivation (Casey and Clarke, 1979; Scholefield et al., 1993). The relationship between soil moisture and nitrogen export from a catchment is complex and depends on the water retaining capacity of the soil, the rates of mineralization and the rates of nitrification and denitrification (Scholefield et al., 1993; Stronge et al., 1997). In a dry summer, low soil moisture levels have a negative impact on bacterial and

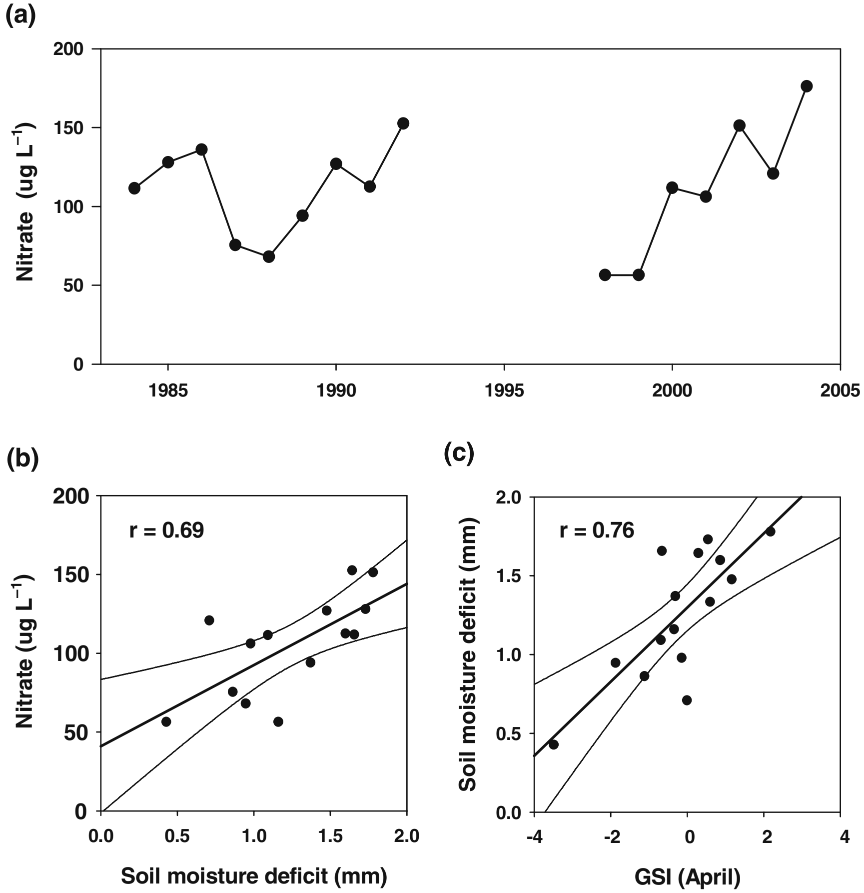

Fig. 19.8 (**a**) The year-to-year variations in the winter concentration of nitrate in Muckross Lake. (**b**) The relationship between the winter concentration of nitrate and the soil moisture deficit recorded the previous summer. (**c**) The relationship between soil moisture deficit in the Muckross area and the Gulf Stream Index (GSI)

plant growth. Positive relationships have consequently been reported between the quantity of nitrate leached in the autumn and the soil moisture deficits recorded during the previous summer (Scholefield et al., 1993; Allott and Jennings, 2006).

In the south west of Ireland, the inter-annual variation in these catchment-based processes have recently been correlated with the north–south movements of the Gulf Stream in the Atlantic (Jennings and Allott, 2006). Muckross Lake in the south west of Ireland is an oligotrophic lake situated in the Killarney National Park. Nitrate concentrations in this lake have not increased over the past twenty years but there are large year-to-year fluctuations in the average winter values (Fig. 19.8a). The analyses reported by Jennings and Allott show that these fluctuations are closely correlated with the position of the Gulf Stream in the western Atlantic the previous spring. The position of the Gulf Stream appears to influence the supply of nitrate by modulating the soil moisture deficits in the surrounding catchment. Figure 19.8b shows the relationship between winter nitrate concentrations in Muckross and the soil moisture deficits recorded the previous summer. The fitted regression explained 48% of the inter-annual variation and was statistically significant at the 99.9% level. Figure 19.8c shows the relationship between these soil moisture deficits and the position of the Gulf Stream in April. The method used to derive the Gulf Stream Index (GSI) has been described by Taylor and Stephens (1980). In this procedure, the latitude of the Gulf Stream's northern limits is abstracted from charts and an index of position constructed using principal components analysis. Positive values of this index are associated with calmer, drier, summers along the west coasts of Britain and Ireland and negative values with windier and wetter conditions. Here, the fitted regression explained 59% of the observed variation in the soil moisture deficit ($r = 0.76$, $p < 0.001$). Interestingly, the correlation between the soil moisture deficit in summer and the April GSI is stronger than that observed between the soil moisture deficits and a range of local meteorological measurements i.e. the index appears to both integrate, and amplify, the imposed climatic 'signal'. Very similar amplification effects have been reported in Britain where the position of the Gulf Stream is known to influence both the dynamics of phytoplankton and zooplankton in the English Lake District (George and Taylor, 1995; George, 2000a) and the relative dominance of herbaceous plants in the south west of England (Willis et al., 1995).

19.6.3 The Biological Effects of Warm, Dry Summers

Most species of planktonic algae can survive and grow at temperatures well in excess of those projected for a warmer world. Hawkes (1969) has examined the temperature tolerance of different groups and suggested that diatoms grow best at temperatures below 25°C and blue-green algae at temperatures above 30°C. There are, however, notable exceptions, such as the diatom *Acnanthes marginulata* which can tolerate temperatures up to 41°C (Patrick, 1969). In physiological terms, most groups of algae photosynthesise most efficiently at temperatures of around 25°C. The rate of carbon fixation might, therefore, increase with increasing temperature, but factors other than temperature usually regulate the seasonal development of phytoplankton. In most lakes, the seasonal succession of phytoplankton is regulated by the combined effects of external factors, like the supply of nutrients, and internal factors, like changes in the frequency and intensity of wind-induced mixing. These physical effects are particularly important for the growth and development of phytoplankton in thermally stratified lakes. Reynolds (1984, 1993) suggests that the seasonal successions observed in these lakes can be explained by the interaction of two processes:

a) An autogenic process, where a number of different 'functional types' dominate the phytoplankton community in an ordered sequence.

b) An allogenic process, where periodic disturbances disrupt this sequence and return the phytoplankton community to a less stable state.

In the lakes of the English Lake District, the principal factor influencing the switch from an autogenic to an allogenic pattern of succession is the intensity of wind-mixing. The first functional group to appear is the diatoms that grow when the water is cold and there is relatively little light. In the spring, when the concentrations of dissolved silica required by this group are depleted, the community becomes dominated by flagellates that grow quickly, but only do well if the water column is periodically mixed by the wind (George and Hewitt, 2006). As the summer advances and there are more prolonged periods of calm, the flagellates are followed by larger, slow-growing forms like dinoflagellates and blue-green algae (George et al., 1990). Figure 19.9a shows the effect that year-to-year variations in the stability of the water column had on the growth of the blue-green alga *Aphanizomenon* in Esthwaite Water between 1960 and 1978. Here, the *Aphanizomenon* numbers have been de-trended to compensate for the progressive enrichment of the lake and the stability of the water column estimated using the procedure described by Schmidt (1928). The results suggest that changes in the stability of the water column play a key role in the seasonal development of this species. There is a strong positive correlation ($r = 0.82$) between the de-trended abundance of *Aphanizomenon* and the stability of the water column, a relationship that was statistically significant at the 98% ($p < 0.02$) level. The main factor regulating the stability of the water column at this sheltered site was the number of anticyclonic days recorded during the summer (Fig. 19.9b). Here, the number of anticyclonic days is based on the Lamb catalogue of daily weather types (Lamb, 1950) now produced by the automated system described by Jenkinson and Collison (1977). The analysis demonstrates that the changing frequency of this weather type accounted for a significant proportion of the historical variation in the mixing characteristics of the lake. The fitted linear regression was statistically

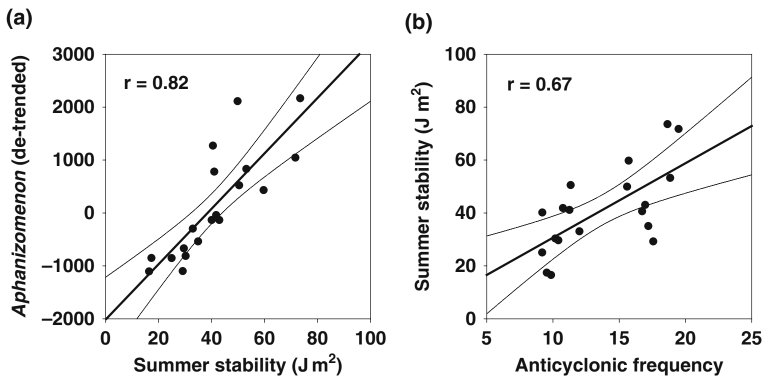

Fig. 19.9 (**a**) The relationship between the de-trended abundance of Aphanizomenon in Esthwaite Water and the stability of the water column. (**b**) The relationship between the stability of the water column and the frequency of calm, anticyclonic days

significant at the 95% ($p < 0.05$) level and explained 45% of the observed variability in the calculated stability.

19.7 The Impact of Extreme Weather Events

19.7.1 The Physical Effects of Climatic Extremes

Most climatologists believe that the frequency of extreme weather events will increase as the world becomes warmer. In Britain and Ireland, Hulme et al. (2002) suggest that there will be a marked increase in the number of very warm summer days i.e. days when the summer temperatures exceed the 90% threshold in current records. By the 2080s, the north of Scotland is projected to experience at least ten ‘very warm’ days every year whilst the total for the south of England may be three times higher.

There has already been a noticeable increase in the maximum summer temperatures recorded in many lakes and reservoirs. In Britain and Ireland, much of this increase can be related to the increased frequency of anticyclonic days but this cannot fully account for the recent increases observed in some of these lakes. Figure 19.10a relates the number of days when the summer surface temperature of the North Basin of Windermere exceeded 20°C to the observed variation in the number of anticyclonic days. In the 1960s, the average frequency of very high surface temperatures was 1.1 days per summer but this increased to 11.2 days in the 1990s. Figure 19.10b shows the number of days when the sur-

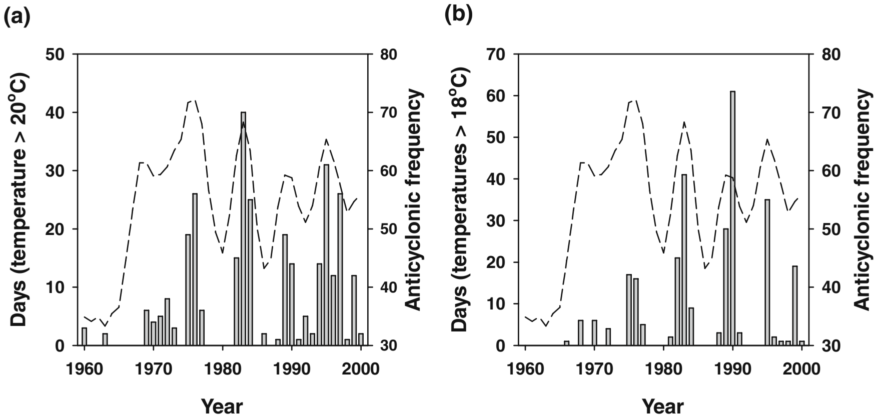

Fig. 19.10 (**a**) The year-to-year variation in the number of days when the surface temperature of the North Basin of Windermere exceeded 20°C and the corresponding variation in the number of anticyclonic days (- - -). (**b**) The year-to-year variation in the number of days when the surface temperature of Lough Feeagh exceeded 18°C and the corresponding variation in the number of anticyclonic days (- - -). The time-series of anticyclonic days has been smoothed using a three-point running mean

face temperature of Lough Feeagh exceeded 18°C. Here, the frequency of relatively high surface temperatures increased from an average of 0.7 days in the 1960s to an average of 12.2 in the 1990s. The 'mismatch' observed between the water temperature and the meteorological record implies that the frequency of these high surface temperatures is already higher than expected from the number of anticyclonic days. In both lakes, low temperatures were recorded in the early 1970s when the number of anticyclonic days was high and high temperatures in the 1990s when the number of anticyclonic days was consistently lower.

19.7.2 The Chemical Effects of Climatic Extremes

In recent years, the colour of the water abstracted from many upland areas in Britain has increased (Watts et al., 2001; Worrall et al., 2004; Evans et al., 2006). There is mounting evidence to suggest that much of this increase is due to the periodic drying and re-wetting of peat (Naden and McDonald, 1989; Mitchell and McDonald, 1992; Clark et al., 2005). The increased production of coloured compounds is thought to be due to the aerobic breakdown of organic matter in dry, warm soils. The products of this breakdown accumulate in the soil and then enter the drainage system when rain follows a prolonged drought (Chapter 4, this volume).

The example in Fig. 19.11, taken from a paper by Watts et al. (2001), shows the long-term change in the colour of the water processed at a treatment works in the north of England. At this site, the highest colour levels were recorded in years that followed the very dry summers of 1989–1990 and 1995–1996. The variations in the colour of the water follow a seasonal cycle with the highest concentrations recorded in early winter when the humic compounds that have accumulated in the soil during the summer are washed into the reservoir. Prolonged periods of drought disrupt this

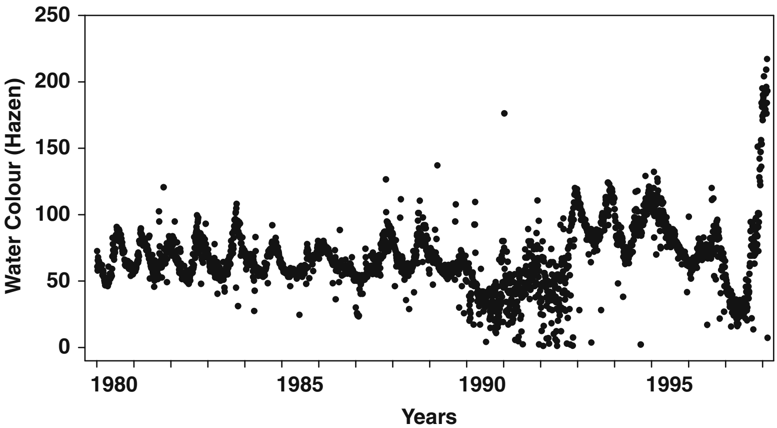

Fig. 19.11 The year-to-year variations in the colour of the water processed at a treatment works in the north of England. (Modified from Watts et al., 2001)

regular cycle and give rise to physical and chemical changes that can last for several years. Dry peat is highly hydrophobic and is very difficult to re-wet. When it starts to rain again, humic acids do not immediately appear in solution since the wetting process can be slow and there may be changes in their solubility. The colour of the drainage water may thus remain low for some time (Watts et al., 2001) but the subsequent increase can be very rapid. In Fig. 19.11 the colour response that followed the drought of 1995–1996 was more extreme than that recorded in the early 1990s. More recent measurements confirm that colour levels at this site have remained high. Analysis of colour data from another site in the north of England (Naden and McDonald, 1989) showed a similar response to soil moisture between 1979 and 1987. Here, an ARIMA transfer function model, based on soil moisture deficits, explained 68% of the variance in the water colour data and there were highly significant correlations at lags of 13 and 14 months.

19.7.3 The Biological Effects of Climatic Extremes

A good example of the effect that an increase in the frequency of calm, sunny days can have on the growth of phytoplankton is that observed in the Queen Elizabeth II (QE II) reservoir in the south east of England. The QE II was built in 1962 to provide additional storage capacity for the city of London. Water from the River Thames is normally stored in the reservoir for a maximum of 40 days before being treated and distributed to customers. Chlorophyll concentrations below 20 $\mu g\ L^{-1}$ present no problems for the water treatment plant but higher concentrations reduce the efficiency of filtration. In the 1960s and 1970s, the quality of the water stored in the reservoir during the summer was very good but problem growths of algae have become increasingly common in the 1990s.

Figure 19.12a shows the year-to-year variation in the concentration of chlorophyll *a* measured in the surface waters of the QE II between January 1980 and December 2000. In the early 1980s, chlorophyll *a* concentrations greater than 20 $\mu g\ L^{-1}$ were rare but concentrations in excess of 40 $\mu g\ L^{-1}$ are now observed almost every year. Nutrient concentrations in this reservoir are so high that they are never limiting but variations in incident light and the mixing characteristics of the reservoir have a profound effect on the development of the phytoplankton (George et al., 2005). Figure 19.12b shows the relationship between the average concentration of chlorophyll measured in the reservoir in late summer and number of hours of bright sunshine recorded at a local meteorological station. The sunshine records are daily averages and the fitted regression explains 38% of the observed variation in the biomass of the phytoplankton. When the QE II was built, provision was made for mixing the reservoir at regular intervals by 'jetting' the water pumped from the River Thames. These procedures are now much less effective due to the increased stability of the water column. A detailed analysis of the pumping records and the meteorological data suggest that the factor responsible for this change was the progressive increase in the number of hours of bright sunshine (Fig. 19.12c). In the early 1980s, the average number of hours of bright sunshine

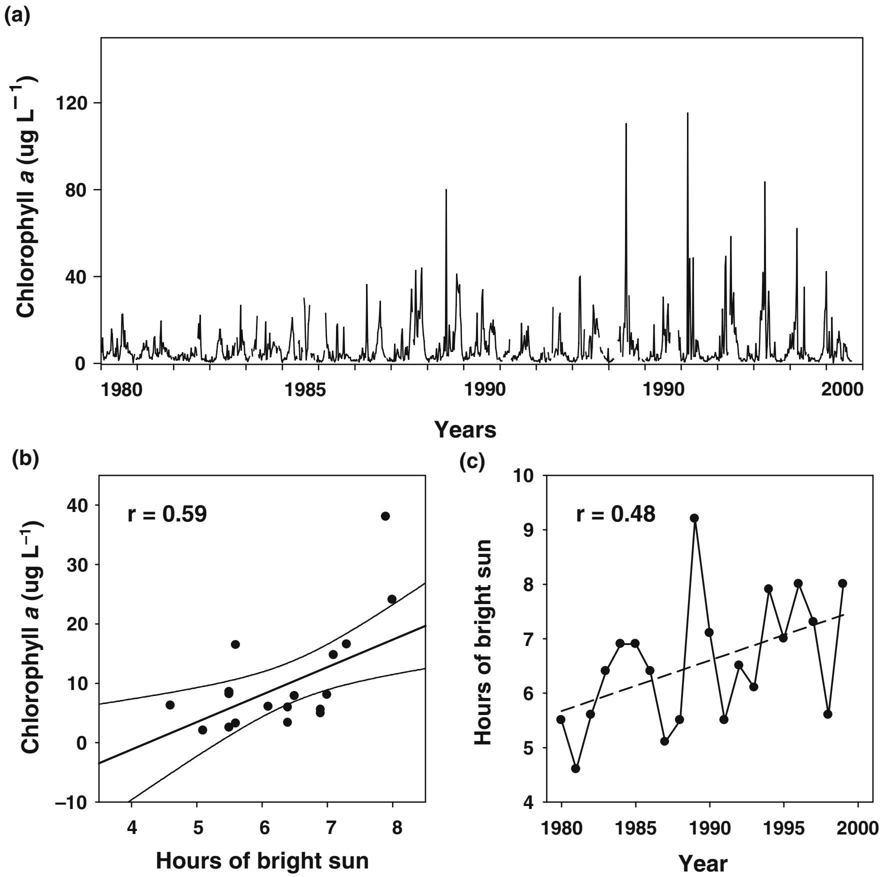

Fig. 19.12 (**a**) The year-to-year variation in the concentration of chlorophyll measured in the surface waters of the QE II reservoir between January 1980 and December 2000. (**b**) The relationship between the concentration of chlorophyll in late summer and number of hours of bright sunshine. (**c**) The long-term change in the number of hours of bright sunshine

was 5.9 but this increased to an average of 7.1 in the late 1990s. The frequency and severity of these summer blooms is likely to increase as the world becomes warmer. New procedures may then have to be devised to enhance the rate of mixing when there is not enough water in the river to run the jets. The reservoir is already fitted with pumps for internal re-circulation but the cost of running these pumps is high and will increase as energy prices rise.

19.8 Discussion

In this chapter, we have described some of the ways in which long-term changes in the weather have influenced the winter and summer dynamics of lakes situated on the west coast of Britain and Ireland. These lakes are known to be very sensitive to the movement of weather systems across the Atlantic (George and Harris, 1985; George and Taylor, 1995; George, 2002), atmospheric features that are likely to change in a warmer world. The examples presented are all based on the analysis of data acquired over several decades. Historical comparisons of this kind provide a very powerful way of quantifying the effects of historical variations in the weather on the seasonal dynamics of lakes. From a climate change point of view, we were fortunate that the period selected included a number of mild winters and very warm summers. It is, however, important to note that the data analysed here were acquired from lakes that are relatively deep and do not support extensive stands of macrophytes. Much less is known about the effects of long-term changes in the weather on the shallower lakes found in southern Britain and the midlands of Ireland. In shallow lakes, the projected changes in the air temperature and the rainfall are likely to have a more pronounced effect on their ecology than the seasonal variation the wind-speed. The heat budgets of these lakes are more directly influenced by the sensible transfer of heat across the air-water interface and even the colour of their bottom sediments can influence the heat content of the water. In these shallow systems, the most important biological effects are likely to be those associated with the growth of macrophytes and the switch between alternate stable states (Jeppesen et al., 1998; Blindow et al., 2006; Scheffer and Jeppesen, 2007). Some indication of the complexity of these responses can be gained from the microcosm experiments described by Moss et al. (2003) and McKee et al. (2003). The results of such manipulations have, however, to be treated with caution since simplified systems of this kind are very susceptible to the 'domino effects' noted by Carpenter (1996).

The historical data analysed here show that winters in Britain and Ireland have become much milder and wetter in recent years. A considerable proportion of this variability can be related to changes in the NAO. This atmospheric feature has been shown to influence the dynamics of lakes throughout Europe (George et al., 2004; Blenckner et al., 2007). In northern Europe, the most important effects are those associated with the extension of the ice-free period and the timing of the spring phytoplankton maxima (Weyhenmeyer et al., 1999; Blenckner and Chen, 2003). Elsewhere, the effects include: changes in the flux of nutrients (Monteith et al., 2000; George, 2000b), the timing of the clear water phase (Straile et al., 2003) and the seasonal abundance of crustacean zooplankton (George, 2000d). Variations in the NAO have a particularly pronounced effect on the weather experienced along the Atlantic coast. Here, there is a strong positive correlation between the winter rainfall and the NAOI but the correlations recorded further inland are more variable (Wilby et al., 1997). One question, still to be resolved, is the extent to which the trend towards positive values of the NAOI is being driven by global warming. There is some evidence to suggest that the recent positive phase of the NAO is linked to climate forcing (Paeth et al., 1999). Recent model simulations by the UK Hadley Centre (Hulme et al., 2002) also imply that there is a functional connection. Experiments based on the 'Medium-High' emissions scenario indicate that this trend is likely to continue with the index becoming significantly different from its historical average sometime around 2050.

The climate change projections for Britain and Ireland (Chapter 2, this volume) suggest that winters in the region will become milder and wetter. In all the scenarios, there was a steep northwest to southeast gradient in the projected changes. The most extreme values, produced by the Max Planck model, imply that winter rainfall could increase by as much as 40% in the north and west and at least 20% in the south and east. The corresponding increases in the winter air temperature ranged from 2.0°C in the north to 3.5°C in the south. Changes of this magnitude will have a significant effect on the lakes by eroding soils, increasing nutrient loads and regulating both the growth and seasonal succession of phytoplankton. In the near future, the most important effects will be those associated with the increased frequency of extreme rainfall events. May et al. (2005) have shown that 90% of the total sediment load to a lough in the west of Ireland was deposited during a series of very short-lived extreme events. This study was based on the high-resolution measurements acquired by an automatic monitoring system which provided the only effective means of measuring the transport of sediment by floods that lasted less than an hour.

The historical variations in the summer weather show that summers in Britain and Ireland have tended to become warmer in recent years. Some of this variation can be related to a systematic change in the dominant circulation types (Chapter 16 in this volume). In summer, the circulation pattern that has the most pronounced effect on the dynamics of the lakes is the high pressure, anticyclonic type. In the period analysed here, the frequency of this weather type has tended to increase and there has been a corresponding reduction in the frequency of westerly days (Hulme and Barrow, 1997). At present, we know very little about the factors responsible for these changes. The analyses produced by Jones and Kelly (1982) and Briffa et al. (1990) have shown that these trends are statistically significant and related to an increased frequency of ‘blocking’ pressure distributions over the Atlantic. More recently, Taylor (1996) has shown that the trajectory of storms in this area is influenced by the north-south movements of the Gulf Stream in the western Atlantic. In southwest Ireland, summers tend to be warmer and drier with lower wind speeds when the Gulf Stream is in a northerly position (Jennings and Allott, 2006). In the English Lake District, early summer wind speeds also tend to be lower when the Gulf Stream moves north and there is corresponding increase in the stability of the lakes (George and Taylor, 1995).

The climate change projections for the summer are even more extreme than those suggested for the winter. For most climatic variables, the steepness of the northwest to southeast gradient, that characterizes the region, will increase. In the southeast, the projections generated by the Max Planck model suggest that average air temperatures could be 5°C higher than they are today whilst the rainfall could be 50% lower. The changes projected for the areas where the main lakes in Britain and Ireland are located are less severe. The more extreme scenarios nevertheless imply that the air temperature could increase by 3.5°C and the rainfall decline by at least 30%. These changes far exceed those experienced during the reference period used in the CLIME studies and would have a profound effect on the ecological status of the lakes. The most important effects will be those associated with the reductions in water level, the increased consumption of oxygen in deep water, the extension of the growing season and the increased frequency of cyanobacterial blooms (George et al., 1990; Paerl and Huiseman, 2008). Most of the lakes considered here are important centres for tourism and recreation. Quite modest reductions in the water level could well have a significant impact on the attractiveness of these lakes and severely limit many water-based sports and pastimes. Elsewhere in this volume, we have presented examples to show that the effects associated with climate change mimic those of cultural eutrophication. These include the enhanced consumption of oxygen in deep water (Chapter 8) and systematic changes in the composition of the phytoplankton (Chapter 14). One issue not addressed in CLIME is the effect that such changes could have on the survival of fish that are already at the limit of their geographical range. A typical example is the survival of the vendace (*Coregonus albula*) in the English Lake District. George et al. (2006) used a 1-D mixing model to simulate the vertical distribution of temperature and oxygen in Bassenthwaite Lake and showed that the recent catastrophic decline in the numbers of this rare species could have been exacerbated by a succession of warm summers.

In this review, we have paid particular attention to water quality variables that are used as diagnostic elements in the Water Framework Directive. The Directive, as currently conceived, treats these variables in a simplistic way and does not take account of the dynamic interactions that play such a key role in the ecology of lakes. There is mounting evidence to suggest that lakes seldom change in a gradual way but switch abruptly from one stable state to another. In shallow lakes, the concept of alternative stable states is well developed (Scheffer, 1990; Scheffer and Jeppeson, 1998) but the thresholds that trigger these shifts are often difficult to define. The critical factor is usually the nutrient loading (Jeppesen et al., 1997) which is, in turn, influenced by the hydrology of the catchment and temperature-related effects in the soil. The regime changes observed in deep lakes tend to be more subtle and typically involve changes in the physical dynamics of the water column. A good example is that described in Chapter 14 for the north basin of Windermere. Here, the extension in the summer growth period for the phytoplankton was correlated with the recent increase in the summer stability of the lake. Future studies on the impact of climate on lakes will have to refine the methods currently used to quantify such non-linear responses. A recent example from the same lake demonstrates that some of these responses are species-specific. When Thackerey et al. (2008) compared the phenology of two diatom species, they found that the spring maximum of one taxon had advanced as a consequence of earlier stratification whilst the dynamics of another was regulated by the combined effects of enrichment and warming.

An important issue, not addressed by CLIME, is the effect that future changes in land-use could have on the ecology of lakes. This question is further complicated by the fact that 'what is grown where' is principally determined by the Common Agricultural Policy. To date, there has been little change in the management of the land in the English Lake District or the south-west of Ireland. This situation could, however, change if upland farming was to be viewed as uneconomic and these areas planted with different crops. For example, the suggestion that more land could be used for coppiced woodland could lead to a significant increase in the amount of water lost by evapotranspiration. It is now clear, that the problems posed by climate change can only be resolved by an integrated approach to lake and catchment management. Water quality and quantity are intimately connected via feedbacks that involve socioeconomic as well as physical and chemical factors. Arnell and Delaney (2006) recently produced a review of the ways in which water companies are adapting to the problem of climate change. Our experiences in CLIME suggest that those responsible for managing water supplies are more aware of these problems than those charged with the care of the natural environment. The Water Framework Directive provides an ideal means of addressing these issues, but more needs to be done to quantify the climatic sensitivity of our lakes at a regional as well as European level.

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Chapter 20
The Impact of Climate Change on Lakes in Central Europe

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20.1 Introduction

In Europe, the effects of global warming are expected to be particularly acute in areas exposed to a more extreme continental climate. The climate change scenarios summarized in Chapter 2, this volume, suggest that the average summer temperatures in some areas of Central Europe could increase by as much as 6°C by 2071–2100. The associated projections for the rainfall give even more cause for concern with the reductions in some areas approaching 50% in summer. In this chapter we analyse impacts of changing weather conditions on lakes in Central Europe. Long-term data sets from a number of lakes are used to link measured variables to climate signals. Particular attention is paid to the lakes in the perialpine region which are known to be very sensitive to short-term changes in the weather (Psenner, 2003; Thompson et al., 2005). Here, the topography and the steep orography enhance the water cycle, and result in flooding, debris flows, avalanches, vertical plant migration etc. The Alps also form a barrier to the mass movement of air and are responsible for the sharp climatic divide between Atlantic, Continental and Mediterranean influences.

Central Europe is a variously and vaguely defined region. Rather than a physical entity, it is more a reflection of a shared history. The results summarized here are based on the analysis of long-term climatological and limnological data from the countries shown in Fig. 20.1. These include Germany (DE), Poland (PL), the Czech Republic (CZ), Slovakia (SK), Switzerland (CH), Lichtenstein (LI), Austria (AT) and Hungary (HU). The Central European countries are geographically diverse with landforms ranging from the North-German Lowlands, through the Alps to the Hungarian plain. The pannonian plain in the eastern part is also a major climatic 'crossing point' and is affected by the Eastern-European continental, the Western-European oceanic and the Mediterranean influence.

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Fig. 20.1 Map of Europe showing the location of the seven Central European countries defined in the text

20.2 The Region and the Lakes

Conceptually, lakes are commonly regarded as 'filters' which integrate and amplify the local climate (Blenckner, 2005; George, 2006). The lakes in Central Europe include a range of very different types, such as riverine lakes, deep stratifying holomictic lakes, meromictic and shallow lakes as well as artificial reservoirs. Lakes in the perialpine Region are mainly of the 'alpine lake' type. Although this term is widely used as a descriptor for a specific lake type, a precise definition is not possible. Depending on their origin and elevation, three main categories of alpine lakes can be distinguished with several sub-types in each category (Dokulil, 2004):

– high alpine lakes – high altitude, above the tree-line,
– alpine lakes – glacial valley or fjord-type lakes,
– pre- or subalpine lakes – at lower elevations in the perialpine lowlands.

The lakes in the northern lowlands range from shallow, polymictic, hypertrophic to deep, dimictic, oligotrophic. Müggelsee represents the shallow riverine lake type (Driescher et al., 1993). Heiligensee and Stechlinsee exemplify shallow and deep stratifying lakes respectively (Gerten and Adrian, 2001). High-altitude mountain lakes are found on both the Slovak and Polish sides of the Tatra Mountains (Gregor and Pacl, 2005). Reservoirs are widespread throughout the region but are especially numerous in the Czech Republic and in Germany (e.g. Bucka, 1998; Desertová and Punčochár, 1998; Horn, 2003).

According to Meybeck (1995), lakes in the Alps cover approximately 3,440 km^2 and small lakes of less than 0.1 km^2 in size are most abundant. The largest lakes in the region are Lac Leman (581 km^2) and Lake Constance (593 km^2). Most lakes are deep and therefore stratify during summer. Freezing during winter largely depends on lake size and elevation (Dokulil, 2004). The mixing regimes of the lakes are either dimictic when they freeze, or warm monomictic if they mix throughout the winter. Lakes at higher elevations are cold-monomictic and ice-covered for the greater part of the year (Eckel, 1955). The main difference between the alpine lakes and lakes in other deglaciated regions is the much greater depth of the lakes. The maximum to mean depth ratio of 0.46 is very similar to the world average and is also close to that commonly reported for deep lakes (Dokulil, 2004). Several lakes in the region are characterised by metalimnetic populations of the filamentous cyanobacterium *Planktothrix rubescens*. Examples include Mondsee, Wolfgangsee in Austria and Ammersee in Bavaria.

In the south, the region includes the largest shallow lake in Central Europe, Lake Balaton (mean depth 3.2 m) and the alkaline, turbid steppe lake Neusiedler See (mean depth 1.2 m), both located on the Austro-Hungarian plain (Padisák, 1998). This area has a marked continental climate with abundant sunshine and high summer temperatures. As a consequence, the lakes are of the polymictic lake type which stratify only occasionally at irregular intervals. Both these lakes were originally endorheic, but now have artificial, regulated outflows. The water budget strongly depends on evaporation. Water temperatures follow changes in air temperatures with a time lag of hours and are very high during summer and often below freezing in winter.

20.3 The Historical Variations in the Weather

The climatological variations that characterize the region are illustrated in Fig. 20.2. In some cases, there is very little differences between the annual averages but the amplitude of the seasonal variations are much greater at the more continental locations.

Near the northern edge of the Alps at Salzburg, Austria precipitation is twice as high as that recorded elsewhere, exceeding 100 mm during the summer months. The average air temperature here is slightly below that at the other two stations. At the high altitude site on the peak of Sonnblick in Austria climatic conditions are entirely different. The annual average air temperature is –5.8°C and exceeds zero degrees for only about two month per year. Precipitation, at over 100 mm in almost every month, accumulates to produce a mean total of 1,832 mm per year.

The average increase in the observed annual mean temperature across Europe for the last century was 0.8°C. During the same period, above surface air temperatures in Austria increased by 1.8°C. The temperatures recorded during the winter have, in general, increased more than those recorded during the summer. In 2003, many countries in central Europe experienced the warmest summer on record and the last 30 years was the warmest period in the last five centuries. The winter of 2006/2007 was also the mildest on record and was followed by an unusually warm spring. Annual precipitation over Northern Europe has increased by between 10 and 40% in the last century while the Mediterranean basin has experienced reductions of up to 20%. In future, summer heat waves and intense winter precipitation events are projected to become more frequent and the risk of drought is likely to increase throughout central and southern Europe.

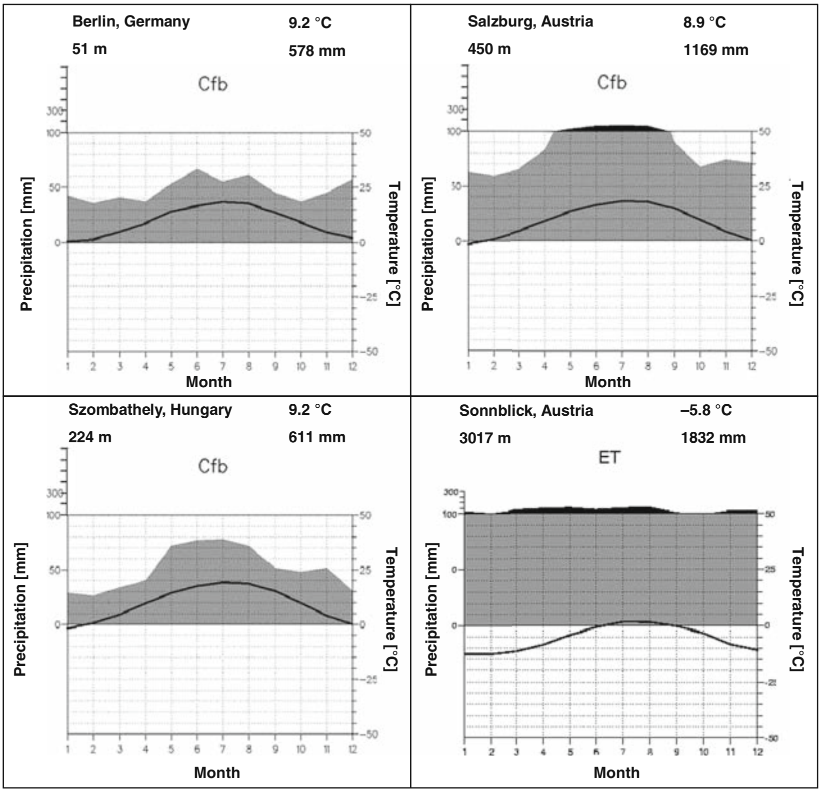

Fig. 20.2 Diagrams showing the climatological variations for three cities and a high elevation site in Central Europe (Berlin, Germany, Szombathely, Hungary, Salzburg and Sonnblick, Austria). Climate classification according to Köppen: Cdf = temperate, humid climate with warm summers; ET = Tundra climate. (Modified from Mühr, 2006)

The temperature increase in the Alpine region are synchronous with the global warming but have much greater amplitude (Fig. 20.3). The rate of warming is twice as great as the global average reaching anomalies close to 1°C and up to 2°C for individual sites (Beniston et al., 1997). Differences between west, east, north, south or sea level are also statistically significant (Auer, 2003; Auer et al., 2006). Long-term precipitation trends are highly variable between seasons, have large spatial differences or are even antagonistic in direction. Long-term trends in the NW and SE have even abruptly changed into their opposite in recent times (Auer, 2003, 2006; Böhm, 2006; Brunetti et al., 2006).

20.4 The Climate Change Projections

Several projections of future changes in climatic conditions are now available but many are based on different scenarios for the change in atmospheric CO_2 concentration. The regional climate model simulations documented by Räisänen

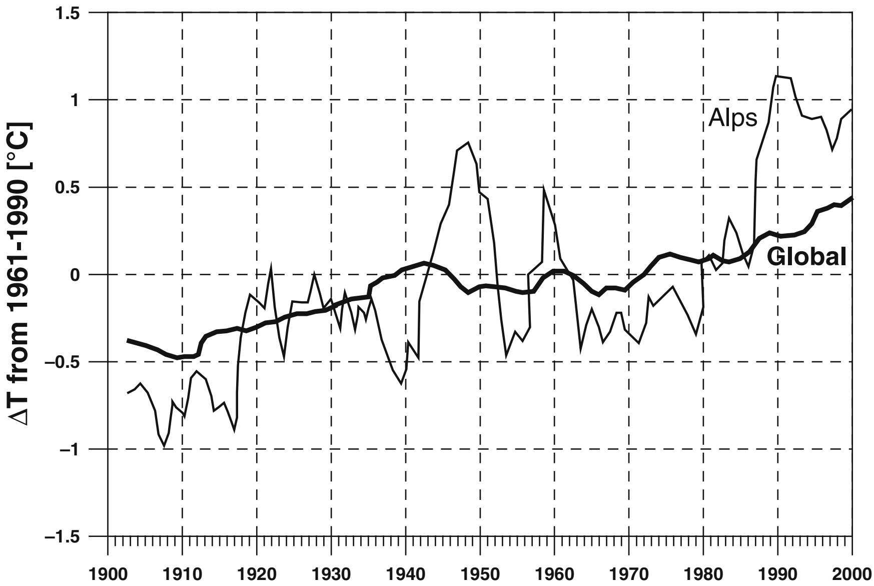

Fig. 20.3 Annual average temperature anomalies from high elevation sites in the Alps compared to global mean temperature anomalies. Expressed as deviations in °C from the IPCC base period 1961–1990 and smoothed with a five-year filter (Modified from Beniston et al., 1997)

et al. (2003) all indicate that warming in central Europe will be greatest in summer, locally reaching 10°C. This increase in temperature is associated with substantial decreases in precipitation, soil moisture and cloudiness. The projected temperature increase for the year 2035 is 2.0–3.5°C depending on the scenario and whether or not the damping by aerosols are considered (Kromp-Kolb and Formayer, 2001). In central Europe, the precipitation is projected to increase by at least 30% in winter and decrease by as much as 50% in summer (Chapter 2, this volume). The magnitude of these variations varies geographically and there is a general tendency for the number of precipitation days to decrease in central and southern Europe. The yearly extremes increase even in the areas of central Europe where there is a decrease in the mean annual precipitation. The simulated annual mean evaporation increases in most of central Europe under all the selected scenarios.

20.5 The Influence of Global-Scale Changes in the Weather

Variations in the climate are usually characterized by 'anomalies' which are the deviations of the short-term situations from the long-term climatic average. Within these global circulation patterns there are some recurrent patterns of variability. For the Northern Hemisphere and Europe, the North Atlantic Oscillation (NAO) is of prime importance (Marshall et al., 2001; Ottersen et al., 2001). Based on this and other pressure gradients, a number of indices have been proposed as indicators for the teleconnections that regulate the climate in specific geographical regions.

The most commonly used proxy for Central Europe is the NAO-index (Hurrell, 1995). The NAO can be characterised by a family of related indices which all follow the same broad pattern of variation with no significant dominant periodicities (Stephenson, 2006). The most closely related index is the Arctic Oscillation (AO) which follows the same winter pattern (Shindell et al., 2001). In the positive phase, it is a better measure of the number and frequency of days with subzero temperatures or substantial snowfall in the mid-latitudes (University Washington, 2001). Another proxy used to explore the climatic responses of lakes situated close to the Atlantic coast is an index based on the position of the north-wall of the Atlantic Gulf stream, GSI (Taylor and Stephens, 1998; Taylor et al., 1998). This index is significantly correlated with the NAO at a lag of two years (George and Hewitt, 1998; Jennings and Allott, 2006) but adds very little to our understanding of weather patterns in Central Europe. Another index useful in more continental situations is the Mediterranean Oscillation index (MOI), defined by Palutikof et al. (1996) and Conte et al. (1989) as the normalized pressure difference between Algiers (36.4°N, 3.1°E) and Cairo (30.1°N, 31.4°E). On a regional scale, indices based on the classification of daily weather types have also proved useful and include those developed by Chen (2000), Lamb (1972) and Steinacker (2000) which are described in Chapter 16 of this volume. A reduced set of the circulation types proposed by Steinacker (2000) for the Alps has recently been used by Nickus and Thies (2004) to explore the effect of inter-annual variations in the weather on alpine lakes. The western, north-western and variable flow type was closely correlated to the winter NAO ($r^2 = 0.42$, $p < 0.001$). Analyses that relate the decadal trend in these regional indices to the NAO show a clear contrast between two recent decades. The years 1961–1970, dominated by negative NAO values, were characterized by the north-weather flow type while 1991–2000 had positive NAO values associated with increased frequencies of the west and variable flow types.

20.5.1 Impacts on Temperature, Stability and Timing of Events

One of the most important physical parameter in any lacustrine system is the temperature of the lake since it reflects meteorological forcing in a direct and sensitive way (Dokulil, 2000). In temperate regions, the highest surface water temperatures in winter are recorded in deep lakes that retain heat and the lowest in shallower lakes that loose more heat to the atmosphere. In ice-covered lakes the most sensitive climatic indicators are usually the timing and duration of ice cover (Chapter 4, this volume) but significant effects have also been observed in the open water (Chapter 6, this volume). At very cold locations, the year-to-year variations in the winter weather can also have important effects on the ecology of the lakes. These include the impact of the physical characteristics of the ice and the type and depth of snow on the underwater light regime and the convective currents that develop under the ice (Chapter 5, this volume). Changes in the freeze-thaw cycle of lakes at high latitudes and altitudes are often used as proxy indicators of regional changes in the weather (see Magnuson et al., 2000 and Chapter 4, this volume).

Long distance climatic forcing also affects the processes that regulate the summer dynamics of the lakes. These include the onset, timing and duration of thermal stratification, the depth and intensity of mixing and the heat content. Lake surface water temperatures (LSWT) are closely correlated with the air temperature at all elevations (Livingstone et al., 2005a). Both variables are strongly influenced by the North Atlantic Oscillation (NAO), especially during winter and spring (Livingstone and Dokulil, 2001). Some example correlations for several lakes, not analysed by Livingstone and Dokulil (2001), are shown in Fig. 20.4. These include Neusiedler See in the eastern part of Austria and two lakes from the Carinthian lake-district south of the Alps. Strong teleconnections are evident in all these examples, with the weakest recorded for lakes, like Altausserer See and Grundlsee in the inner alpine valleys, and those located to the south of the alpine ridge. At the more isolated locations, local climatic conditions are far more influential and other weather patterns are of greater importance at sites situated south of the alpine divide.

The situation of Neusiedler See in the pannonian plain suggests that 'Mediterranean' influences could be important there. We therefore compared the temperature records with the Mediterranean Oscillation Index (MOI), an index commonly used to analyse weather patterns in the Mediterranean region (Martin-Vide and Lopez-Bustins, 2006; Maheras and Kutiel, 1999). In fact, the MOI is a better predictor of

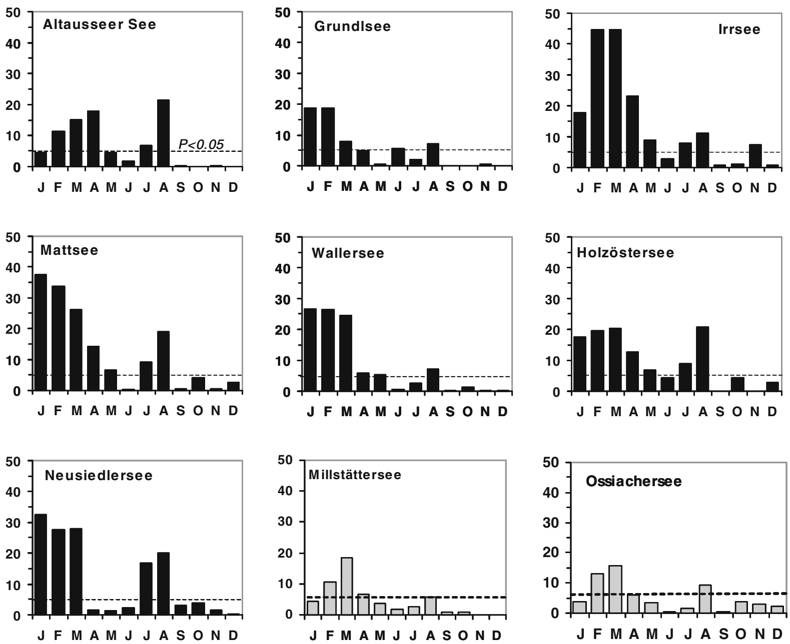

Fig. 20.4 Bar graphs showing the seasonal variation in the coefficients of determination (r^2) between winter NAO index and monthly mean lake surface temperature for nine lakes in the peri-alpine region of Austria

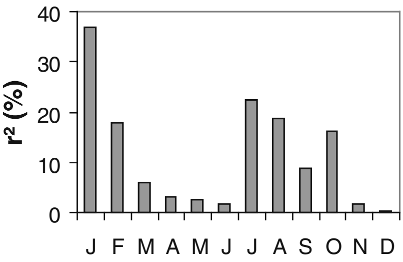

Fig. 20.5 Coefficient of determination (r^2) between winter MOI_{DJFM} index and monthly mean lake surface temperature in Neusiedler See, Austria

summer and early autumn water temperatures in Neusiedler See (Fig. 20.5) but the NAO still has a significant effect on the LSWT in winter.

Long-term trends in LSWT indicate considerable differences between seasons and regions. The average decadal increase between 1940 and 2000 in sixteen lakes north of the Alps was 0.17°C for the winter season and 0.25°C for both spring and summer (Table 20.1). The average rate of change in the autumn was 0.07°C per decade but there were large differences between the individual lakes (– 0.31 to + 0.21). The average increase in LSWT in three middle sized lakes south of the Alps was very similar to that north of the Alps but absolute values were smaller and less variable at all seasons (Table 20.1). The average increase per decade in shallow Neusiedler See was well above the average for the other lakes. Most noticeably, LWST in winter increased on average by 0.31°C per decade, which is the highest rate of change in the whole set of data. In all the lakes and in all seasons except winter, there was a logarithmic relationship between these rates of change and the maximum depth of the lakes. Temperature changes during the winter seem to depend more on orography and ice-cover than the depth of the lake.

The long-term increase in the summer lake water temperature is not simply a reflection of changes in air temperature. As shown by Wilhelm et al. (2006), the rate of increase of the daily night time minimum exceeds that of the daily maximum. This day-night asymmetry in epilimnetic temperatures is influenced by the flux of heat across the air-water interface which, in turn, depends on wind speed, relative humidity and cloud cover. Summer warming will cause increases in the frequency and the duration of stratification events in polymictic lakes such as Müggelsee. Summer stratification events in the extremely warm summers of 2003 and 2006 were the longest on record lasting for up to 8 weeks (Wilhelm and Adrian, 2007). In the long run, lakes currently dimictic or polymictic may well become monomictic in the future, if they are then free of ice during winter.

In the very deep Italian lakes south of the Alps Ambrosetti et al. (2003) found a reduction of winter mixing over the last 40 years and, consequently, the deeper layers have become less affected by seasonal variations which retain a 'climatic memory' in their hypolimnia. In such deep lakes, complete circulation becomes increasingly difficult affecting lake hydrodynamics, turnover, deep water chemistry and dissolved gases. Mixing events then require more energy, that is more wind,

Table 20.1 Long-term increase in lake surface water temperature (LSWT) for the four seasons

Lake	Geogr. position N/E	Elevation [m a.s.l.]	Surface area [km^2]	Max depth [m]	Season Winter	Spring	Summer	Autumn
Bodensee	47°38′/9°22′	395	536.00	254.0	0.3066	0.2395	0.2524	0.0927
Piburger See	47°11′/10°53′	913	0.13	25.0	0.2099	0.3478	0.2166	0.2273
Zeller See	47°19′/12°48′	750	4.55	68.0	**0.3989**	0.1641	–0.0655	0.1113
Holzöster See	48°03′/12°54′	460	1.82	4.7	0.0434	**0.7616**	**0.7854**	0.1299
Mattsee	47°59′/13°07′	503	3.60	42.0	0.3054	0.1690	0.3672	**0.2097**
Wallersee	47°54′/13°10′	506	6.10	23.0	0.1248	0.1036	–0.0426	–0.0004
Fuschlsee	47°48′/13°16′	636	2.65	66.3	0.0845	0.1605	0.3546	0.2076
Irrsee	47°54′/13°18′	533	3.47	32.0	0.0667	0.7222	0.2257	–0.3099
Mondsee	47°49′/13°22′	481	14.21	68.3	0.2212	0.3462	0.3278	0.1742
Attersee	47°54′/13°33′	469	45.90	170.6	0.1481	0.2038	0.2186	0.0764
Altausseer	47°38′/13°47′	712	2.10	52.8	0.2905	0.2744	0.4990	0.0837
Grundlsee	47°38′/13°51′	709	4.14	63.8	0.0715	0.0472	0.1131	–0.0316
Hallstätter See	47°34′/13°39′	508	8.58	125.2	0.0137	0.0903	0.1276	–0.1354
Wolfgang S.	47°45′/13°23′	538	12.84	113.1	0.0764	0.0861	0.2071	0.0727
Traunsee	47°52′/13°48′	422	25.60	191.0	0.1339	0.1640	0.1312	–0.0214
Lunzer See	47°51′/15°03′	608	0.68	33.7	0.1625	0.1658	0.2262	0.2120
Mean					**0.17**	**0.25**	**0.25**	**0.07**
Maximum					**0.40**	**0.76**	**0.79**	**0.23**
Minimum					**0.01**	**0.05**	**–0.07**	**–0.31**
Range					**0.39**	**0.71**	**0.85**	**0.54**
Millstätter	46°47′/13°34′	588	13.28	141.0	0.0699	0.1512	0.1497	–0.0228
Ossiacher	46°40′/13°57′	501	10.79	52.0	0.1513	0.0979	0.1980	0.1411
Wörther	46°37′/14°09′	439	19.38	85.2	0.1029	0.1916	0.0269	–0.0414
Mean					**0.11**	**0.15**	**0.12**	**0.03**
Maximum					**0.15**	**0.19**	**0.20**	**0.14**
Minimum					**0.07**	**0.10**	**0.03**	**–0.04**
Range					**0.08**	**0.09**	**0.17**	**0.18**
Neusiedler See	47°49′/16°44′	115	321.00	1.8	0.2444	0.5129	0.6622	**0.3059**

winter = DJF, spring = MAM, summer = JJA, autumn = SON. Data in °C per decade. Lakes are arranged according to geographical position or drainage basin marked by dashed lines. The elevation, surface area and maximum depth for each lake is given

to re-establish initial thermal conditions (Ambrosetti and Barbanti, 1999). The increased surface temperatures and enhanced thermal stability in the summer will affect nutrient flux and the growth of plankton organisms. The nature of these biological effects will largely depend on the physical characteristics and the maximum depth of the individual lake.

20.5.2 Regional Coherence

Despite the large distances separating sub-regions within Central Europe and the different character of the lakes, variations in surface water temperatures were highly synchronous (coherent) and related to fluctuations in air temperature. The surface temperatures in all the selected lakes, ranging from Müggelsee in the north to Lake Balaton in the south, were particularly coherent in summer and autumn. The proportion of variance shared pair wise between the residual time-series always exceeded 30%. For lakes located within tens of kilometres of each other, shared variances of 80–90% was not uncommon (Dokulil and Teubner, 2003; Livingstone et al., 2005b).

In general, coherence between lakes cascades down from physical parameters via chemical and nutrient variables to biological entities (Fig. 20.6 and Dokulil and Teubner, 2002). In other words, the effects of changes in climatic conditions on the biology of lakes are complex, difficult to disentangle from other influences, and not easy to generalise (see Livingstone et al., 2007 and Chapter 17, this volume).

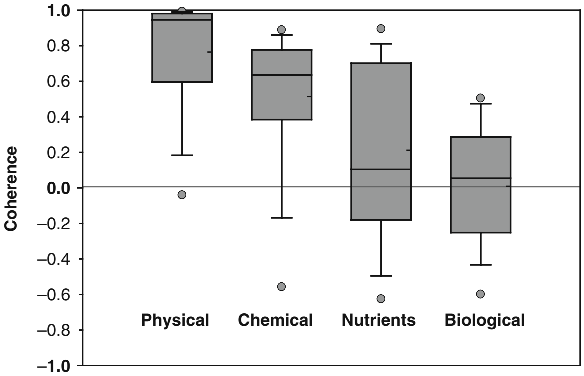

Fig. 20.6 Regional coherence (expressed as correlation coefficients) between pairs of six alpine lakes in the Austrian 'Salzkammergut' region shown as box-whisker plots. Box limits are the 25th and 75th percentile; whiskers indicate the 10th and 90th percentile. In these boxes, the *solid line* is the median, the *dashed line* the mean. Physical = surface temperature, light attenuation and Secchi-depth; chemical = pH, conductivity and oxygen concentration; nutrients = total phosphorus, total nitrogen and dissolved silica; biological = chlorophyll-a and phytoplankton biomass. (Modified from Dokulil and Teubner, 2002)

20.5.3 Chemical and Biological Effects

Small variations in climate are likely to have a particular dramatic effect on extreme habitats, such as high altitude lakes (Psenner, 2003) or shallow soda lakes (Kirschner et al., 2002). In these systems, many species live at the limit of their capabilities and will respond very quickly to any additional stresses such as a change in the duration of ice cover or summer drought. If these habitats dry out or become fragmented many species will disappear and be replaced by species with special adaptations. The effects of global warming on phytoplankton dynamics are not fundamentally different in different regions of the world (Gerten and Adrian, 2002b). In winter, effects are connected with light conditions, ice duration and variations in wind-induced mixing in ice-free lakes. Winter conditions are also largely responsible for the timing, magnitude and composition of the spring peak of phytoplankton, the clear water phase and the development of the zooplankton in both shallow and deep lakes (Gerten and Adrian, 2000; Straile, 2000). The propagation and cascading effect of the NAO on meteorological conditions, water temperature, zooplankton biomass and on the clear water phase has been discussed by (Straile 2000, Straile et al., 2003b) and is shown schematically in Fig. 20.7. These results demonstrate that the timing of the spring phytoplankton bloom, as measured by chlorophyll-a is strongly related to climate forcing during the winter.

In Mondsee (Fig. 20.8), the strongest correlation was that observed with an index derived by averaging the January and February NAO index of Jones et al. (1997). Here, the average timing of the spring chlorophyll maximum has advanced by about 48 days. In Northern Europe, the spring bloom of phytoplankton is more closely correlated with the NAO index for March (Weyhenmeyer et al., 1999). Simulation of phytoplankton growth during winter and spring in Upper Lake Constance, a deep monomictic lake, also indicates that changes in the meteorological conditions associated with warming alter the onset of phytoplankton growth and subsequent succession of species (Peeters et al., 2007). The phytoplankton growth rate increases when these lakes start to stratify but is soon regulated by zooplankton grazing, particularly from ciliates.

In summer, prolonged thermal stratification can influence hypolimnetic oxygen conditions, dissolved nutrient concentration and phytoplankton composition (Jankowski et al., 2006; Wilhelm and Adrian, 2008). Oxygen depletion and higher temperatures increase nutrient release processes at the sediment-water interface (Søndergaard et al., 2003) and increase the stress on aquatic organisms (Weider and Lampert, 1985; Saeger et al., 2000; Wilhelm and Adrian, 2007).

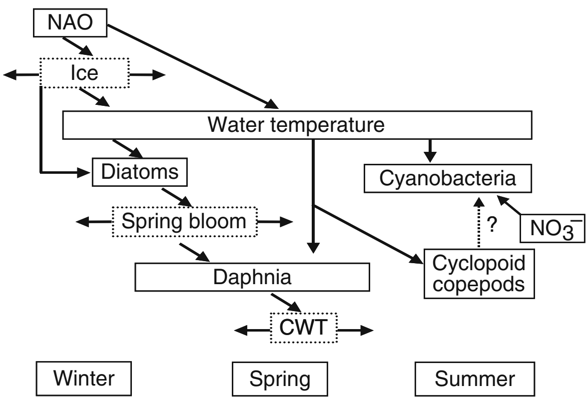

Fig. 20.7 Conceptual diagram of the effects of Winter NAO on lake ecosystems. Modified from Blenckner et al. (2007) after Straile (2000)

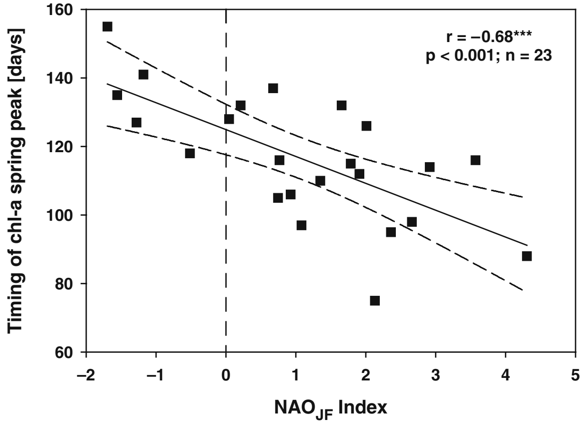

Fig. 20.8 The relationship between the timing of the chlorophyll-a spring peak in Mondsee (1982–2004) and the average January–February NAO Index (NAO_{JF})

An analysis of 81 shallow lakes from three climatic regions including lakes from Central Europe suggests that climate is an important predictor of zooplankton biomass, community composition and food-web dynamics (Gyllström et al., 2005; Straile, 2005). In Lake Washington, USA the trophic linkage between phytoplankton and zooplankton is disrupted by the systems differential sensitivity to vernal warming. In many lakes, the long-term decline in *Daphnia* populations is associated with the temporal mismatches that develop with the spring diatom bloom (e.g. Winder and Schindler, 2004). The response of freshwater copepods to the recent increase in the summer temperature appear to be species specific (Gerten and Adrian, 2002a).

In the shallow eutrophic Müggelsee, warming has increased both the number and size of newly developed larvae of *Dreissena polymorpha*, suggesting that conditions for overall reproductive success have improved. A sudden drop in the abundance of the larvae in 2003 is attributed by Wilhelm and Adrian (2008) to the low dissolved oxygen concentrations associated with an unusually long period of stratification.

The potential impacts of climate on fish are summarised by Ficke et al. (2005). A good example from Central Europe is that observed in Lake Constance (Straile et al., 2007). In this lake, the temperature variations associated with NAO affects the year-class strength of fish (*Coregonus lavaretus*, Blaufelchen) by regulating both the egg development time and larval growth rate. The duration of egg development is related to the NAO with a time lag of one year due to the mixing characteristics of this warm, monomictic lake. Larvae hatch earlier if the previous winter has been relatively warm.

20.5.4 Teleconnections

The long-term increase in temperature associated with mild winters influences the timing of ice formation, ice duration and ice-break up date. Both duration and ice-off are significantly correlated with winter NAO in lakes of Central Europe

Table 20.2 Correlation coefficients of ice conditions in Central European lakes versus winter NAO, AO and MOI. Lakes are arranged from North to South

Lake	Ice	NAO	AO	MOI
Müggelsee	Duration	–0.762***	–0.612***	n.s.
	Ice-off	–0.609***	–0.504*	n.s.
Irrsee	Duration	–0.494***	–0.410***	n.s.
	Ice-off	–0.671***	–0.330*	n.s.
Mondsee	Duration	–0.570**	–0.443*	n.s.
	Ice-off	–0.724**	–0.774**	n.s.
Neusiedler see	Duration	–0.451*	n.s.	–0.503*
	Ice-off	–0.511**	–0.461*	–0.650**
Balaton	Duration	–0.261*	n.s.	–0.381*
	Ice-off	–0.528***	–0.323**	–0.486**

*** = $p<0.001$, ** = $p<0.01$, * = $p<0.05$, n.s. = not significant

(Table 20.2). This teleconnection to climate signals is quite robust and does not depend on the version of the NAO index used (NAO_W Hurrell, 1995; NAO_W Jones et al., 1997). Ice conditions in all lakes are also significantly related to the AO also known as the Northern Annular Mode (NAM), a further indication of long distance climate effects. Lakes of continental position in Central Europe, e.g. Neusiedler See or Balaton (VITUKI, 1996) are also influenced by another, completely different, weather situation. Here, ice duration and ice-out is also related to the Mediterranean Oscillation (MOI) which is not the case in the other lakes listed in Table 20.2.

In common with other lakes in the Perialpine Region north and south of the Alps (Blanc et al., 1990; Livingstone, 1993, 1997; Ambrosetti and Brabanti, 1999), deep water warming was also evident in several lakes across Europe (Dokulil et al., 2006). Hypolimnetic temperatures increased consistently in all lakes by about 0.1–0.2°C per decade. The observed increase was related to large-scale climatic processes over the Atlantic. To be effective, the climatic signal from the North Atlantic Oscillation (NAO) must affect deep lakes in spring before the onset of thermal stratification. The most consistent predictor of hypolimnetic temperature is the mean NAO index for January–May ($NAO_{J–M}$), which explains 22–63% of the inter-annual variation in deepwater temperature in 10 of the 12 lakes (Fig. 20.9). A time lag of one year, as described by Straile et al. (2003a), was not evident but can result from reduced winter cooling and the persistence of small temperature gradients that resist complete mixing. Mixing in turn can determine the trophic status of lakes (Salmaso et al., 2003; Salmaso, 2005) which could have important implications for the assessment of ecological status over time relative to the Water Framework Directive (Eisenreich et al., 2005).

The most important physical process in the annual cycle of many lakes is thermal stratification. The enhanced physical stability of a thermally stratified lake has a profound effect on its chemical and biological characteristic in summer. The direct effects of the projected increases in the temperature may thus be more pronounced in shallow, isothermal lakes but the indirect effects associated with change in

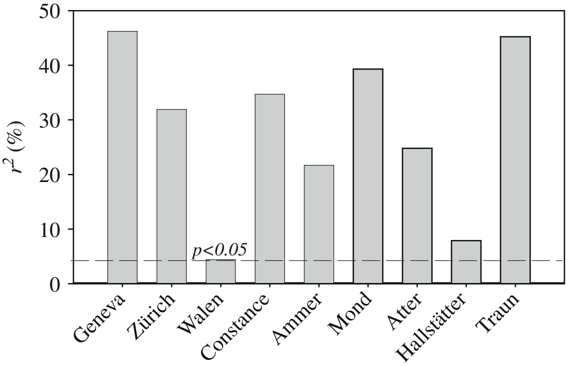

Fig. 20.9 Coefficients of determination (r^2) between the mean NAO index for January–May ($NAO_{J–M}$) and the deep-water temperatures of 9 Central European lakes. Modified from Dokulil et al. (2006)

stability is very important in moderately deep lakes. The responses of thermally stratified lakes to changes in the flux of heat and the intensity of wind mixing are complex and strongly influenced by the morphometry of their basins. The effects of winter NAO are, typically, of short duration in polymictic lakes but in deep lakes with stable summer stratification these signals can even be preserved until the following winter (Gerten and Adrian, 2001). This in turn may even affect the timing and duration of thermal stability in moderately deep alpine lakes. For example, in Lake Mondsee, Austria, (Fig. 20.10) the timing of the onset and maximum thermal stability critically depends on the status of the NAO between January and March. Positive NAO values during the spring period shift the onset earlier while maximum stability is reached later in the year. In contrast, negative values prolong mixing, resulting in later onset and earlier maximum stability.

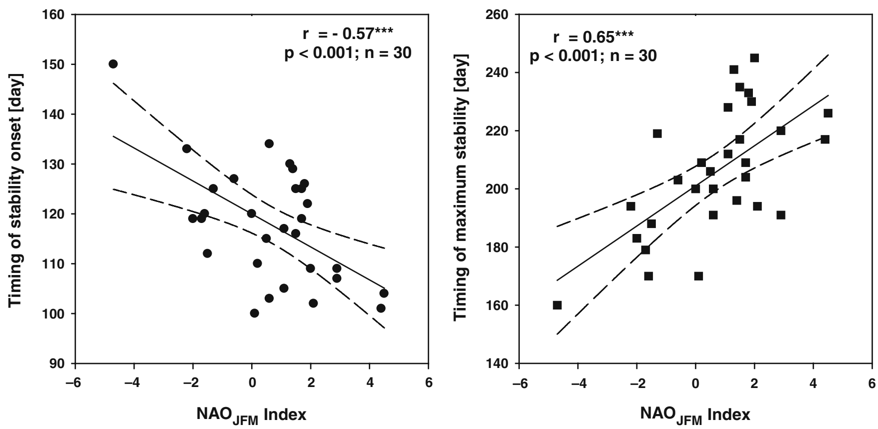

Fig. 20.10 The timing of thermal stratification (*left panel*) and the period of maximum stability (*right panel*) in Mondsee, Austria related to the average January to March NAO of Jones, 1997 (NAO_{JFM})

20.5.5 Impacts from the Catchment

In mild winters, more precipitation falls as rain rather than snow, the time of snow melt is earlier and there is an increase in the peak winter runoff which also occurs earlier in the year. In alpine areas, the accumulation and depletion of snow-pack is particularly important. Changes in the runoff characteristics are expected to be more pronounced in the southern catchments, whilst north facing catchments should be influenced to a lesser extent (Kunstmann et al., 2004).

Changes in nutrient loading to lakes will largely depend on water supply and water chemistry. Phosphorus loading is likely to increase due to enhanced wash-out from soil as a consequence of the increase in net precipitation and the increased frequency of extreme rainfall events (Kromp-Kolb and Schwarzl, 2003). These events are also likely to increase soil erosion ultimately leading to enhanced lake productivity and increased sediment accumulation rate. This will be especially pronounced and important in artificial reservoirs. Nitrogen loading to downstream lakes will be lowered because increased temperatures will result in higher denitrification rates, the main process responsible for N-losses within catchments. The response of in-lake nutrient concentration to climate variability depends on the size of the lake since smaller lakes are more sensitive than large lakes (Jankowski et al., 2005; Straile et al., 2003).

Catchments with a large number of lakes such as e.g. the Elbe river basin (>500 lakes larger than 50 ha) have a major impact on the rivers. If the nutrient loads to these rivers decrease as a consequence of climate change the mixing characteristics and the ecology of these lakes would also change, e.g. there could be a switch from a polimictic to a dimictic state causing drastic changes in the nutrient cycles and the dynamics of the entire ecosystem (Bergfeld et al., 2005).

20.5.6 Extreme Events

As the world becomes warmer the frequency of extreme climatic conditions such as exceptionally mild winters, extremely warm summers, heavy precipitation and flooding or storm events is also expected to increase with major consequences for lakes (Easterling et al., 2000).

In Vienna, the frequency of very warm summers has increased significantly over the last 50 years and there has been a corresponding reduction in the number of very cold winters (Brunetti et al., 2006). Heatwaves in summer are increasingly characterized by large departures from their average values, for example ΔT_{max} was +6.0°C in Basel in August 2003 and reached a maximum of +11.2°C on August 4, 2003 (Beniston and Diaz, 2004). The probability of extreme precipitation events has also increased by 12%, a feature associated with enhanced storm-track activity and 'wetter' conditions over much of central Europe (Palmer and Räisänen, 2002). In Austria, these heavy precipitation events have recently been linked to seven different synoptic patterns (Seibert et al., 2007). Large parts of Central Europe experienced extremely heavy rainfalls during the early days of August 2002 which, in some cases, exceeded the average August totals (DWD, 2002). Extremes of precipitation will also lead to a reduction in the duration of the snow cover period in the Austrian Alps ranging from about 4 weeks in winter to 6 weeks in spring (Hantel et al., 2000). Results for extreme storm events over Central Europe are rather inconsistent. In general, wind speeds tend to increase in winter and decrease in summer (Jonas et al., 2005).

Effects on lake water temperature, stratification and stability differ considerably between extremely cold and warm winters. During cold winters, the effects depend on whether the lake freezes. Frozen lakes have relatively stable water columns but mixing can be intense in open water. Similarly, frozen ground will result in reduced drainage and lower than normal winter nutrient loadings. During warm winters, water temperatures are higher than average and may remain high until spring. Higher than usual spring temperatures can, in turn, lead to higher hypolimnetic temperatures in summer. In Austria, the autumn of 2006 was the warmest on record and was followed by the mildest winter since measurements began. Lakes which usually freeze, such as the shallow Neusiedler See, were not frozen and several deep lakes also remained open throughout the winter.

In summer, very high epilimnetic temperatures are recorded in very warm years. During the 2003 heat-wave the surface temperature of Mondsee exceeded 25°C. The resulting increase in thermal stability can lead to increased rates of oxygen consumption in the hypolimnion which can become anoxic in its deeper parts. This has recently been observed in lake Mondsee, Austria and in a number of Swiss lakes (Jankowski et al., 2006). This increased and extended thermal stability also restricts the vertical transport of nutrients and reduces nutrient concentrations in the epilimnion which, in turn, effects both the composition and the seasonal dynamics of the phytoplankton. Between 2000 and 2003, a series of very dry years at Lake Balaton in Hungary had a major effect on the water budget and the biomass of phytoplankton (Padisák et al., 2006). Less than normal precipitation and increased evaporation drastically decreased lake level. As a consequence, there was an increased incidence of nuisance algal blooms.

Extreme precipitation events increase the loading of nutrients and suspended particles, as was the case with the flood experienced in the Salzkammergut region in 2002. In Mondsee, this increase in turbidity was associated with increased nutrient concentrations, particularly reactive silica (Dokulil and Teubner, 2005). Similarly, Lake Constance experienced dramatic lake level fluctuation during the 1999 centennial flood (Jöhnk et al., 2004).

Short but severe storm events can even have an effect on the water quality of large, shallow lakes. Data from Müggelsee demonstrate that the impact of storms is critically dependent on the antecedent conditions i.e. whether they occur suddenly or build up gradually (Wilhelm et al., 2006).

20.6 Summary and Conclusions

The climate of Central Europe is essentially dependent on weather conditions experienced over the Atlantic. Long-term changes in the circulation of the atmosphere are primarily responsible for the observed changes in the weather. These changes can be expressed by a number of climatic indices. Here we have shown that a clear teleconnection exists between climate signals, weather conditions and the response of the lakes over very large distances.

The rate of warming in Central Europe is now likely to exceed the 0.2°C for the next two decades projected for a range of emission scenarios by the IPCC summary report (2007). Regional differences in Europe are also expected to increase with flash floods and erosion becoming more common. Reduced snow cover and glacier retreat will affect tourism and reduce the quantity of water available to consumers. Central European projections for the summer period suggest drastically increased air temperature and decreasing precipitation. In contrast, winter run-off will significantly increase due to reduced snow cover. The number of extreme events affecting the catchment of lakes is also likely to increase.

Ice cover of lakes is expected to shorten in duration. Lake surface temperature during summer will rise by about 4°C. Deep water temperatures are projected to increase by about 0.1–0.2°C per decade and will have a significant effect on oxygen levels and the internal re-cycling of nutrients. The duration and strength of thermal stratification will expand. Associated with these changes are shifts in algae, plankton and fish abundance.

Increased erosion due to high winter run-off combined with higher water temperatures and more prolonged stratification in summer will, almost certainly, lead to widespread, climate-related eutrophication.

Alterations in catchment processes will affect the external loading of nutrients and change the residence time of most lakes. Together with in-lake changes, particularly in nutrient cycles which are biologically mediated and thus temperature dependent, climate change will adversely affect the objectives of good water quality as defined in the Water Framework Directive (WFD). The anticipated changes depend strongly on lake type and local conditions. Biological changes induced by climatic changes are inherently unpredictable because of complex interactions which influence shifts in community structure and increase the risk of harmful algal blooms, the severity of deep water anoxia and the subsequent mortality of commercial fish species. In addition to these anthropogenic changes, there is also natural variability. To enable distinction, the WFD introduced a detailed typology of water bodies as the basis for water quality assessment. As the world becomes warmer, these typologies will need to be revised and new reference conditions established to take account of the changing climatic conditions (see Chapter 24, this volume).

Monitoring fresh-water ecosystems in a changing world requires not only permanent meteorological and hydrological observations, but also a network of automatic water quality stations to record the short-term responses of lakes and rivers. Such stations should, ideally, be installed at sites where long-term environmental data is already available. For large water-bodies, such observations can readily be augmented by remote sensing. Only then we will be able to answer the questions how much natural variation can be accommodated within types and how to differentiate between natural variation and impact. To cope with the more pronounced, irreversible impacts, the classification scheme of the WFD will need to be updated at regular intervals with due regard to the non-static reference conditions that characterize a warmer world.

Finally, socio-economic development and changes in socio-economic structures in combination with climate change must be considered which could seriously alter natural conditions by e.g. habitat destruction and loss of biodiversity. Socio-economic impacts of climate trends will clearly not be the same in all the European regions. Some areas might benefit while others are adversely affected. Overall, climate change will certainly affect many aspects and facets of Central Europe and poses a constant challenge for future development.

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Chapter 21
Developing a Decision Support System for Assessing the Impact of Climate Change on Lakes

Ari Jolma, Teemu Kokkonen, Harri Koivusalo, Hanne Laine, and Külli Tiits

21.1 Introduction

CLIME was a complex project where data from a variety of sources had to be summarized in a form that was readily accessible to stakeholders. The volume of data generated by the regional climate models, the catchment models, and the lake models was also very large and placed severe constraints on their direct examination. In situations such as this, Decision Support Systems (DSS), which employ databases, visualization, and models in an interactive way are usually the tools of choice (e.g., Loucks and da Costa, 1991; MacEachren et al., 2004). The Internet and the Web are increasingly being used as platforms for DSS, as systems implemented in this way are easy to access and to maintain (Jolma, 1999; Jolma et al., 1999). The analysis of causalities and the development of probabilistic models are issues that are central to any projections linked to climate change. Varis (1997), Ames (2002), Borsuk et al. (2004), and others have obtained encouraging results by applying Bayesian networks to environmental problems.

In CLIME, the key relationships analyzed were those between the water quality variables and the changes projected by the Regional Climate Models (RCMs). Our central objective was to produce a computer-based tool that would allow the stakeholders to explore the effect of these changes on the dynamics of lakes situated anywhere in Northern, Western and Central Europe. The key challenge in developing the DSS was the transfer of knowledge from the research domain, via the engineering domain into the management and decision-making domain. The CLIME-DSS was specifically designed to provide strategic support for the implementation of the European Water Framework Directive (WFD): – see Chapters 1 and 23, this volume. The objective of the WFD is the achievement of 'good status' for all European surface and ground waters by the year 2015. In the WFD, the 'status' is defined as a deviation from a specified reference condition. Since these conditions will change as

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the world becomes warmer, managers need a system that can quantify the climatic response of lakes situated in different climatic regions.

21.1.1 Decision Support Systems

Decision Support Systems (DSS) are usually configured as easy-to-use, computer-based systems that allow non-specialists to work with data and models to elicit information that helps the decision making process (e.g., Turgeon and Aronson, 2001). The methods used in a typical data-driven DSS include data warehousing, data mining, case-based reasoning and visualization. In a model-driven DSS, models capture the physical reality and/or the decision maker's preferences in a series of abstractions. Models are used both to achieve optimal or near-optimal solutions to management problems (e.g., McCown, 2002), and to help decision makers select the 'best' choice among a number of alternatives (von Winterfeldt and Edwards, 1986). The simplest models are mathematical equations or functions. Equations including time as a variable can be solved in a sequential fashion, thus setting up a simulation in time, i.e., a simulation model. Other types of models include heuristic models, often based on formal rules defined by experts, and probabilistic models, based on statistical relationships established between random variables. A DSS using a heuristic model is often called an Expert System. Probabilistic models include Bayesian Networks (e.g. Pearl, 2000; Neapolitan, 2003) and time-series models (Box and Jenkins, 1976).

Since DSS is software, the principles of software engineering apply, and the 'end-user' requirements must be established before architecture and functionality of the system is defined. For further details, see Jackson (2001) for the principles of requirements analysis and Nielsen (1993) for the principles of usability. Cognitive science and decision theory also have a part to play in the development of effective, user-friendly decision support system since they explain how people behave and how they can best be supported in their decision-making (see, e.g., French and Geldermann, 2005).

21.1.2 The Objectives of the CLIME-DSS

The development of the CLIME-DSS was a challenging task. It was designed to assimilate large volumes of data from a range of sources, summarize causal relationships, and display the results in both the spatial and temporal domain. From the outset, a decision was made to use the Web to make the results available to as large an audience as possible. The key spatial datasets in the CLIME-DSS were the RCM outputs (Chapter 2, this volume) and the CORINE land cover classification developed by the European Environmental Agency (EEA)[1]. The RCM datasets

[1] http://reports.eea.europa.eu/COR0-landcover/en (accessed 23.3.2008)

were converted from the model-specific geographical grid into the ETRS - Lambert Azimuthal Equal-Area coordinate system used by CORINE, for the computations and visualizations. Climatologists in CLIME provided the RCM data as daily, monthly, and seasonal NetCDF[2] files. The CORINE data were obtained as a 100 × 100 m^2 raster GeoTIFF[3] archive from EEA.

In this chapter, we describe the basic features of the CLIME-DSS and explain how the system can be used to quantify the impact of the projected changes in the climate on the characteristics of lakes situated in Northern, Western and Central Europe. The chapter includes a brief introduction to Bayesian Networks (BN) and some examples to illustrate their application in CLIME.

21.2 Analysis of Projected Changes in the Climate

In CLIME, the change in climate was defined as the difference between the conditions experienced during the reference period (1961–1990) and a specified period in the future (2071–2100). The projections were based on the greenhouse gas emission scenarios developed by the Intergovernmental Panel on Climate Change (IPCC). In CLIME, the researchers only considered the projections based on the A2 and B2 scenarios (Chapter 2 this volume) generated by two different RCMs.

The Graphical User Interface (GUI) of the CLIME-DSS (Fig. 21.1) provides the user with four tools for analyzing the projected changes in the climate: (A) maps showing the projected changes in the climate, (B) tables summarizing the changes expected at particular locations, (C) graphs showing the seasonal variations, and (D) cartographic symbols that show the underlying pattern of climate 'migration'. The tools are controlled by selecting the *climate model*, the emission *scenario*, the climatic *variable*, the *season*, and the type of *map* used for the analysis in the check boxes and menus. Table 21.1 lists the various options that the user can select for these variables. Six climatic variables are available from the RCMs and the emission scenarios are the A2 and B2 scenarios described above. The system summarizes the observed and the projected changes for four seasons: winter (December–February), spring (March–May), summer (June–August), and autumn (September–November). The type of change in the variable is either absolute or relative (see the explanation below). Further details of the driving RCMs are given in Chapter 2, here we explain the functional characteristics of the completed DSS.

The climate maps in the GUI (Fig. 21.1A) were produced from the RCM data by computing the change (difference) between average seasonal values, either as an absolute value, i.e., in mm/season or °C, or as relative value, i.e., percentages. We decided not to display the relative change for air temperature and in some southern areas, where the precipitation was very low, the rate of change for precipitation was set to zero. The maps were coloured in different shades of blue to indicate a decrease

[2]http://www.unidata.ucar.edu/software/netcdf/ (accessed 23.3.2008)
[3]http://www.remotesensing.org/geotiff/geotiff.html (accessed 23.3.2008)

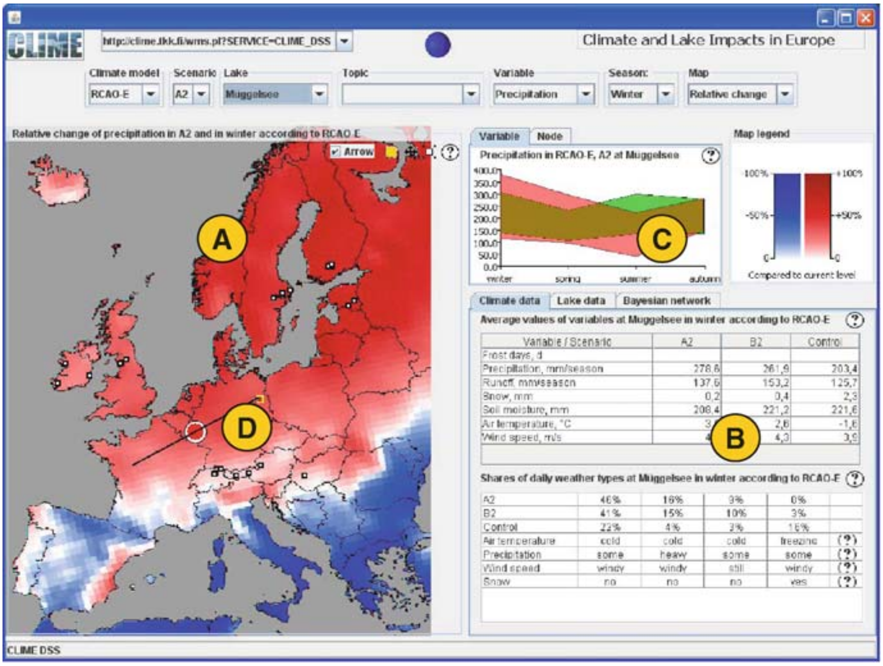

Fig. 21.1 The Graphical User Interface (GUI) used in the CLIME-DSS. The superimposed yellow discs refer to the tools discussed in the text

Table 21.1 The selectable climatic attributes available in the CLIME-DSS. Further details of the regional climate models are given in Chapter 2 (this volume)

Emission Scenario	Climatic variable	Season	Type of change (Map)	Regional climate model
A2	Precipitation	Winter	Absolute	RCAO-E
B2	Runoff	Spring	Relative	RCAO-H
	Soil moisture	Summer		HadRM3P
	Air temperature	Autumn		
	Wind speed			
	Snow			

and different shades of red to indicate an increase in any selected variable. At maximum change, the saturation was set at 100% for each absolute or relative variable.

Two different tables in the GUI show the change in the climate projected at the location selected by the user (Fig. 21.1B). The tables are based on two datasets that were computed from the RCM simulations and are stored in the database of the CLIME-DSS. The datasets were: (i) the average values of the climatic variables, and (ii) average number of days having a given weather type. The values were computed and stored for all emission scenarios, for all seasons, for all RCMs, and for all RCM grid cells. The weather type on a particular day was defined by a simple combination

Table 21.2 Classification of the climate variables used to describe the day types

Class	Precipitation	Runoff	Snow	Soil moisture	Air temperature	Wind speed
1	None (0–0.1 cm/d)	None (0−0.1 cm/d)	None 0	Dry < 210 mm	Freezing < 0°C	Still < 3 m/s
2	Some (0.1−5 cm/d)	Some (0.1−4 cm/d)	Snow > 0	Wet ≥ 210 mm	Cold (0−10°C)	Windy ≥ 3 m/s
3	Heavy (5−∞ cm/d)	Heavy (4−∞ cm/d)			Warm (10−20°C)	
4					Hot ≥ 20°C	

of words (linguistic values). For example the word 'cold' was used to describe a day whose daily average temperature is between 0 and 10°C. The type of day is defined according to all available meteorological variables. For example, a day type can be 'none, some, none, dry, cold, still', where the values are the linguistic values for precipitation, runoff, snow, soil moisture, air temperature, and wind, respectively (see Table 21.2).

A graph in the GUI (Fig. 21.1C) shows the seasonal variation in the selected variable under the future as well as the current conditions. The graph uses three colors: green to denote the current variation, red to denote the projected variation, and brown to denote the region where these variations overlap. The seasonal variations are shown by the minimum and maximum values generated by the RCMs for the selected variable.

The patterns of climate change at any location can also be visualized by the climate 'migration' option (Fig. 21.1D). This tool displays the location in where the current climate at another European location most resembles the projected climate for the selected location. The location pairs are pre-computed and stored in the database. The resemblance between the two climates was determined by a comparison of weather patterns (Jolma et al., 2005). This method was developed in CLIME and uses vectors, whose values are the frequencies of the different day types. The day types are defined using the linguistic variables described above and the discretization of all the values listed gives 288 different day types. The climate at a location is thus expressed as a vector, which has 288 elements, each representing the probability of a particular day type at the location. The similarity of two climates can then be computed as the Euclidean distance between two vectors.

21.3 Analysis of Climate Change Impacts on Lakes

21.3.1 Bayesian Networks (BN) and their Use in the CLIME-DSS

In the CLIME-DSS, Bayesian networks (BN) were used to quantify the climatic effects recorded at various sites as well as those simulated by the lake and the catchment models. A BN is a graphical representation of the behavior of a system as a set of variables (nodes) connected by probabilistic dependencies (links). The variables of a BN are anything that the modeler can attach a probability or belief to. The variables and their values can also be used to represent decisions and utilities. The links in a BN are *directed*, i.e., a link connects a *parent* node to a *child* node. Traditionally, the links represent direct statistical dependence, but they can also be used to represent cause-effect relationships (*from* a cause *to* an effect). Efficient algorithms have been developed for the purpose of reasoning with a BN, i.e., propagating the effect of change in the beliefs attached to one node to the other nodes. See Pearl (2000) for an authoritative discussion and Neapolitan (2003) for some good illustrations. The main limitation of Bayesian networks is that they cannot be used to explicitly model feedback, i.e., the directional graphs must be acyclic. The CLIME-DSS uses the BNJ (Bayesian Networks in Java[4], and Joehanes (2003)) software library for the management of BNs and their computation in the GUI application.

The variables in a BN are usually discretized and the links between each child node and its parents are expressed as conditional probability tables (CPTs) (Fig. 21.2 and Table 21.3). The CPT represents the strength of the relationship between a child and its parents. This relationship can sometimes be weak (characterized by a flat frequency distributions) or strong (characterized by frequency distributions with a strong tendency to certain values). Note that these discretized values represent all possible states for a variable. Whilst a variable can only be in one of these states at any particular time, sequential or spatially distributed measurements yield frequency distributions.

The variables that populate the BNs in the CLIME-DSS are seasonal or yearly averages. The year-to-year variation of these averages makes the variables prob-

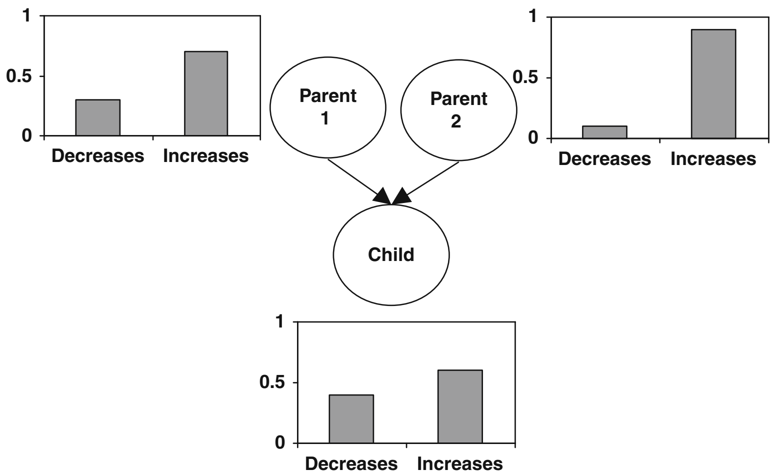

Fig. 21.2 A schematic representation of a Bayesian network with the state distributions of two parent nodes and one child node shown as histograms

[4] http://bnj.sourceforge.net/ (accessed 6.7.2008)

Table 21.3 The Conditional Probability Table (CPT) for the network in Fig. 21.2. The CPT is a matrix of probabilities (p)

		Child	
Parent 1	Parent 2	Decreases	Increases
Decreases	Decreases	$p^{Decreases}_{Decreases,Decreases}$	$p^{Increases}_{Decreases,Decreases}$
Increases	Decreases	$p^{Decreases}_{Increases,Decreases}$	$p^{Increases}_{Increases,Decreases}$
Decreases	Decreases	$p^{Decreases}_{Decreases,Increases}$	$p^{Increases}_{Decreases,Increases}$
Increases	Decreases	$p^{Decreases}_{Increases,Increases}$	$p^{Increases}_{Increases,Increases}$

abilistic. In order to obtain variables that are not site-specific, a reference value was subtracted from the values of the time series before computing the frequency distribution. The average from the reference period (1961–1990) was used as the reference value and the derived relative variables are described as 'deviation' variables in the final network. The expected value of the deviation is zero in the reference period, but non-zero in the projected state affected by the climate change. The main assumption behind the BNs is that the CPTs, which link the 'deviation' nodes, capture the relationship between parent and child variables in such a way that BNs can be applied on geographically large areas under different climatic conditions. The BNs vary in this respect somewhat depending on how the CPTs were estimated. Some CPTs were estimated based on empirical equations while other CPTs were based on running simulation models for both current and in future conditions. In the first case the historical data was typically acquired from a range of lakes, but the observations were obviously limited to the current climate. In the latter case the model was calibrated against measurements acquired at a few locations, and the models then perturbed by both the current and projected climates. When the deviation derived by the BN is added to a user-specified reference value (in our case the average from reference period), the network generates the projected value (average value plus climate change impact).

21.3.2 Development of Bayesian Networks for the CLIME-DSS

The structure of a BN represents the qualitative knowledge collated for the system being modeled. The quantitative knowledge of the system is embedded in the CPTs which, in CLIME, were obtained by running the simulation models. The CLIME-DSS uses BNs to relate the changes in the physical, chemical, and biological variables of lakes to the projected changes in the climatic variables. In practice, the structure of the individual BNs was based on a combination of empirical equations, simulation models, and expert advice. The final, version of the CLIME-DSS included BNs for: (i) timing of lake ice-off, (ii) lake surface temperature, (iii) lake mixing depth, (iv) lake residence time, (v) nitrogen load, (vi) phosphorus load, (vii) dissolved organic carbon load, and (viii) lake chlorophyll concentration. In the following sections we describe the methods used to construct these networks and provide some examples of their application at different locations.

Table 21.4 The Bayesian networks developed for the CLIME-DSS and the models used for their parameterization

Bayesian network	Model	Site	Reference
Timing of ice-off	Empirical equation	Swedish lakes	Weyhenmeyer et al. (2004)
Surface water temperature	Empirical equation	Lakes in the English Lake District	George et al. (2007)
Mixing depth	PROBE	Erken	Chapter 7 this volume
Lake residence time	Empirical equation	Lakes in the English Lake District	George and Hurley (2003)
Nitrogen load	GWLF	Esthwaite, Mustajoki	Chapter 11 this volume
Phosphorous load	GWLF	Esthwaite, Mustajoki	Chapter 9 this volume
DOC concentration	GWLF	Lough Feeagh, Lough Leane, Moor House, Mustajoki	Chapter 13 this volume
Growth of cyanobacteria	PROTECH	Esthwaite	Reynolds et al. (2001)

A key issue in the parameterization of Bayesian Networks is the increasing size of the CPTs as the discretizations of variables in the nodes become denser and/or the number of parent nodes increases. This problem is especially marked where the CPTs are based on expert knowledge but can also exist when simulation model results are used. Use of simulation model results is problematic when all combinations of all states of the parent nodes do not appear in the simulation data. In this case the corresponding probabilities of the child node being in any state (a row in the example in Table 21.3) cannot be estimated.

To minimize these problems, we used a computation scheme that is based on link strengths in parameterizations of the BNs. Our scheme is based on earlier work by Varis and Kuikka (1994) and Srinivas (1993) and has been described in detail by Kokkonen et al. (2005). In our scheme, a single link strength parameter that has a value between –1 and 1 is attached to each link. A link strength that has the value of one characterizes a perfect one-to-one relationship between two nodes, and a link strength that has the value of zero indicates independent nodes. A negative link strength value indicates that there is a negative relationship between the two nodes.

Table 21.4 lists the models that were used to parameterize the BNs in the CLIME-DSS and the key references of the models. The complexity of the models varied from simple empirical equations to complex, process based simulation models. Very large datasets had to be assimilated to develop the BNs that were based on complex models. For example, to parameterize the DOC BNs the GWLF-DOC simulation model was run using more than 50,000 years of daily meteorological values generated by the stochastic weather generator described in Chapter 2.

21.3.3 The Bayesian Network for the Timing of ICE-OFF

The Bayesian network for the timing of ice-off was based on the empirical model described by Weyhenmeyer et al. (2004). They combined data from 196 Swedish

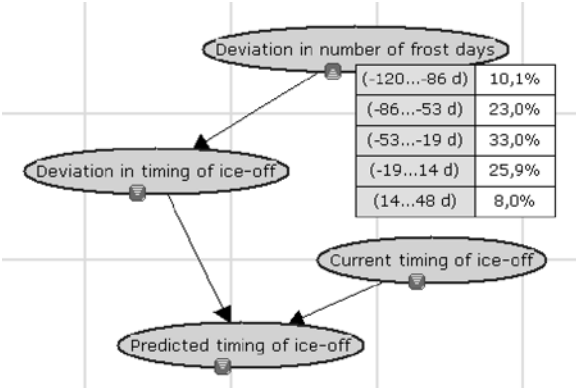

Fig. 21.3 The Bayesian network for the timing of ice-off. The probability distribution for the deviation in number of frost days is stored in an associated table, which is filled by location specific data in the GUI of the CLIME-DSS (the figure shows an example).

lakes covering 13° of latitude and a minimum period of 28 years (Chapter 4 this volume), and related the timing of ice-off to the number of frost days. For the BN, we computed the spatial distribution of the collated deviation of the number of frost days in Europe. Based on the data and the model of Weyhenmeyer et al. (2004) we estimated a CPT that relates to probability distributions of the deviation of the timing of ice-off to the deviations in the number of frost days. The lowest node of the network in Fig. 21.3 is used to add the thus obtained projection for the deviation to the absolute value of current timing of ice-off to obtain a projection for the timing of ice-off in the selected 'warm world' scenario.

21.3.4 The Bayesian Network for the Surface Water Temperature

The Bayesian networks for the winter and summer variations in the lake surface temperatures (Fig. 21.4) were based on the empirical model described by George et al. (2007). In winter, this model calculates the surface temperature of the lakes from the observed or projected air temperatures and from the known or assumed mean depth of each lake. In summer, the values used are the observed and projected air temperatures and the known or assumed depth of the seasonal thermocline. The original equations were based on 40 years of weekly measurements from four different lakes in the English Lake District. In the network we used the deviations recorded during the reference period (1961–1990) as the baseline and assumed that the model would still apply in a warmer world.

21.3.5 The Bayesian Network for Estimating the Mixing Depth of a Lake

The Bayesian network for producing projections for the mixing depths of lakes (Fig. 21.5) was constructed by examining the outputs generated by the one-dimensional physical model PROBE (Svensson, 1998) for a series of very different

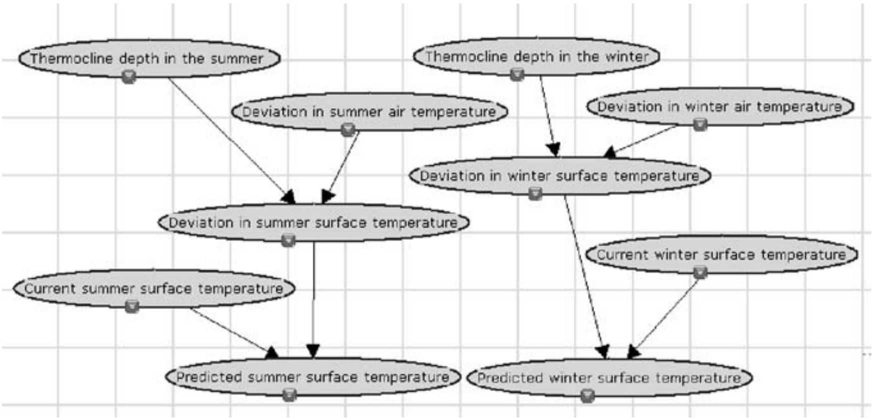

Fig. 21.4 The Bayesian network for the summer and winter temperature of a lake

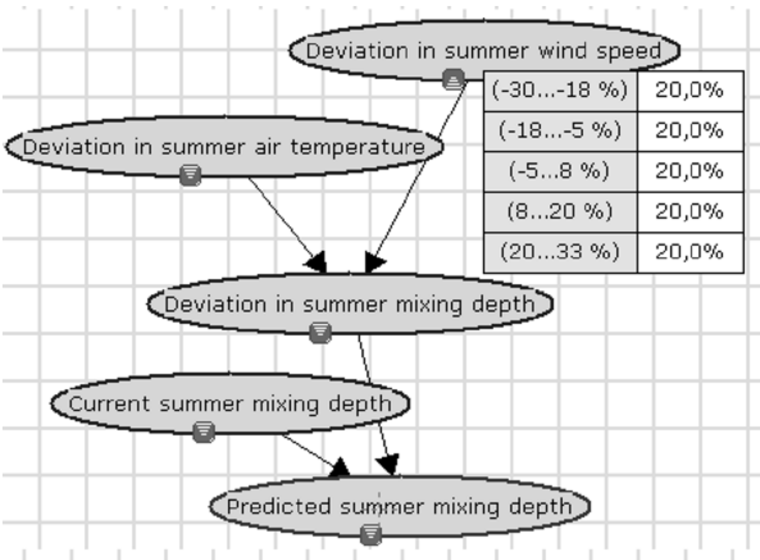

Fig. 21.5 The Bayesian network for the summer mixing depths of a lake. The table shows the probability distribution for the deviation in average summer wind speed

CLIME sites. PROBE simulates the vertical variations in the temperature of a lake using hourly estimates of air temperature, humidity, wind velocity and solar radiation. A detailed description of the model is given in Chapter 7 (this volume) whilst Chapter 15 presents some site-specific applications.

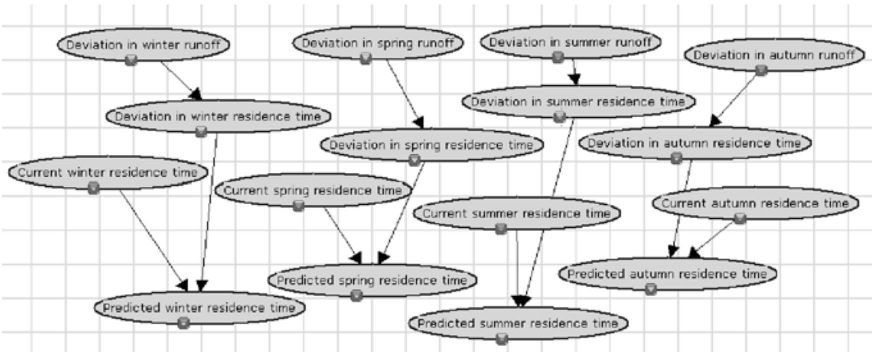

Fig. 21.6 The Bayesian networks for the seasonal variation in the lake residence times

21.3.6 The Bayesian Network for Estimating the Residence Time of a Lake

The Bayesian networks for estimating the residence times of the lakes (Fig. 21.6) were based on the procedure described by George and Hurley (2003) and applied to fifteen lakes in the English Lakes by George et al. (2007). This procedure calculates 'instantaneous' values of the residence time by relating the seasonal variations in runoff to the effective volume of each lake. In winter, the effective volume of each lake is estimated from its known or assumed mean depth. In summer, the effective volume of the isothermal lakes was also calculated from their mean depth but that of the thermally stratified lakes was assumed to be the volume contained in the epilimnion. In the network, we again used the average of recorded values during the 'historical' reference period (1961–1990) as the baseline and assumed that the procedure captured into the CPTs of the BNs still applies in a warmer world.

21.3.7 The Bayesian Networks for the Supply of Phosphorus and Nitrogen

The Bayesian networks for estimating the seasonal flux of nitrogen and phosphorus to the lakes (Fig. 21.7) were based on data generated by the GWLF catchment model. This model combines the results produced by the RCMs with monthly estimates of the phosphate-phosphorus and inorganic nitrogen loads in catchments with a mix of different land uses (Haith and Tubbs, 1981; Haith and Shoemaker, 1987). Further details of the hydrological component of the GWLF model are given in Chapter 3 and case studies on the phosphorus and nitrogen simulations in Chapters 9 and 11. GWLF is a hybrid model that combines a dynamic representation of the hydrological processes in a catchment with an export coefficient approach to the estimation of nutrient fluxes. GWLF is driven by daily temperature and precipitation data and computes the daily water balance, streamflow, runoff, and groundwater

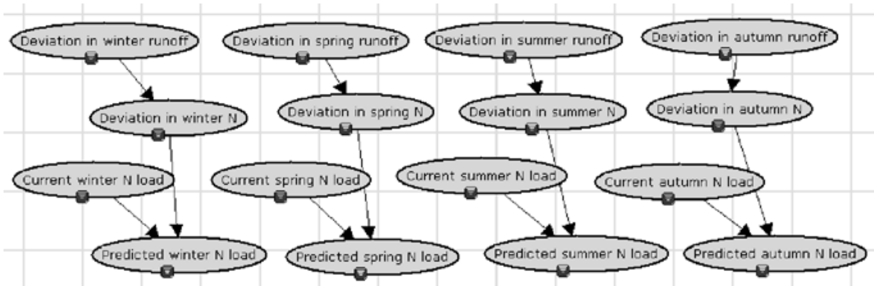

Fig. 21.7 The Bayesian networks for the seasonal variation in the nitrogen loads to a lake. The networks for the phosphorus loads are similar

discharge. The relationships between the deviations in seasonal runoff and the deviations in seasonal nutrient loads were estimated from the model output and used for the CPTs.

21.3.8 The Bayesian Network for Quantifying the Change in the Doc Concentration

The Bayesian network for estimating the project change in the concentration of dissolved organic carbon (DOC) was based on a combination of expert knowledge elicited by a questionnaire and the outputs from the simulation model (GWLF-DOC) described in Chapter 13. The procedure is described in detail by Koivusalo et al. (2005). The final structure of the network is shown in Fig. 21.8. The same

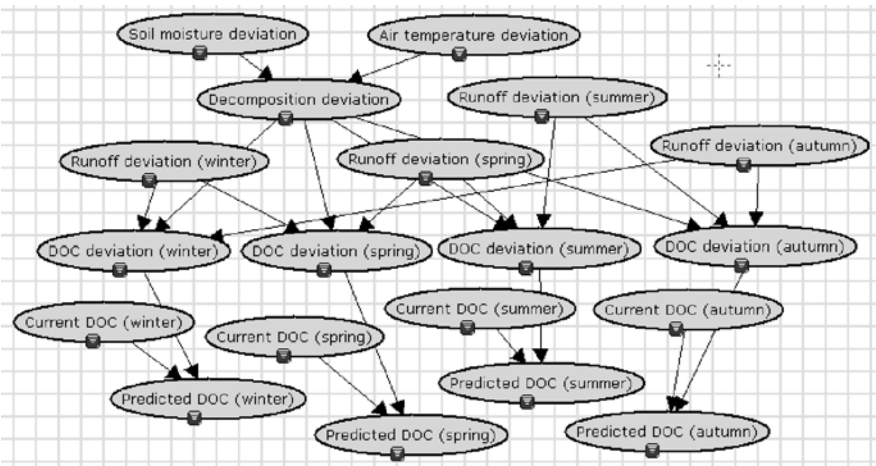

Fig. 21.8 The Bayesian network for the projected deviation in the in-lake concentration of DOC. The soil moisture, air temperature and decomposition deviations are annual values

structure was used for BNs used to simulate the responses expected in the four study areas: Lough Feeagh and Lough Leane in Ireland, Moor-House in the UK and Mustajoki in Finland. In each case the model was calibrated with site-specific data before being perturbed by the climate scenarios described in Chapter 2. The 'link strength estimates' procedure described above was then applied to produce the CPTs.

The main output of the network is the deviation in concentration and is expressed as the discharge weighted concentration in the inflows to the lake. The user needs to provide the current concentration t the point of interest (as a distribution) a, and the BN then adds the concentration deviation to this value to generate the projections for the future conditions.

21.3.9 The Bayesian Network for the Growth of Cyanobacteria

The Bayesian network used to assess the risk of late summer blooms of cyanobacteria (Fig. 21.9) was based on the PROTECH model described by Reynolds et al. (2001). PROTECH is a model that simulates the growth and succession of different functional groups of phytoplankton. The key elements in the model are the equations that define the daily change in the biomass (expressed as chlorophyll *a*) for a representative selection of algal species. Like the other model-based BNs, the CPT estimates in this BN were based on the results of simulations conducted at a site where 'problem' blooms of cyanobacteria were know to occur. The key links in this network were the relationships established between the seasonal variations in chlorophyll *a* and cyanobacteria concentrations and the associated variations in the air, temperature, precipitation, wind speed, depth of mixing, flushing rate and phosphorus load.

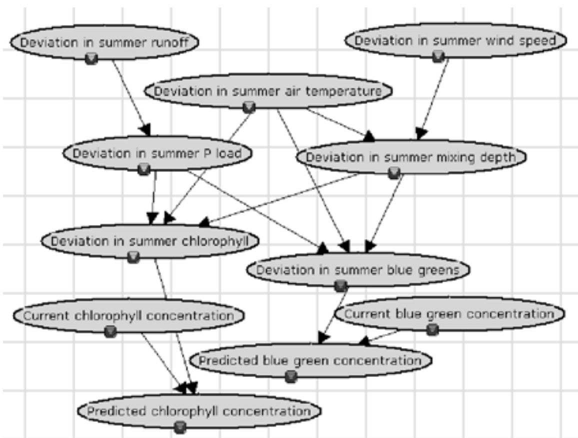

Fig. 21.9 The Bayesian network for the projected deviations in the in-lake concentration of chlorophyll a in the summer and in the amount of blue green algae. Similar networks for other seasons were also developed

21.4 The Architecture of the CLIME-DSS

All the data, models, maps, and visualizations in the CLIME-DSS are centrally managed and relayed over the Internet to the CLIME-DSS GUI application run by the users. Figure 21.10 illustrates the architecture of the CLIME-DSS. Specific server programs provide data to the client applications. The data is managed in a database and files and includes the climate change scenarios, land-cover data, the lakes and their physical characteristics, maps, BNs, etc. The CLIME-DSS GUI application is a Java program that can be downloaded from the CLIME website. It has a built-in capability that allows the users interact with the climate and other data and the BNs.

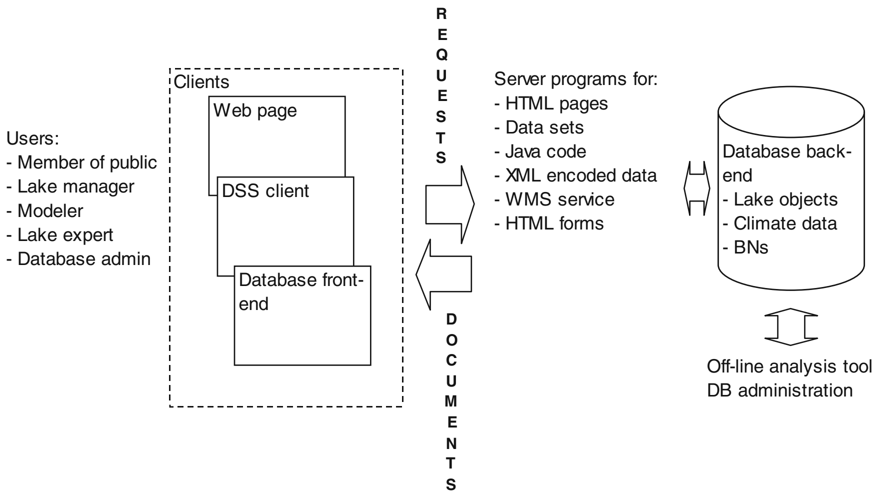

Fig. 21.10 The architecture of the CLIME-DSS, showing the linkages between the Web pages, the GUI, and the database (DB) interface

The system administrators manage the CLIME-DSS using tools provided by the operating system and by the database management software. A typical task is to feed in information created by an off-line analysis tool. The database and model administrators use the web-based interface to manage both the database and the model base. A typical task is to upload and add a BN into the system.

A detailed description of the CLIME database and the server architecture are beyond the scope of this chapter. The key features of the architecture are:

- The database has object-oriented capabilities, and new objects such as lakes, can be added through the Web-based interface.
- The server includes GIS functionality and key attributes such as regional land-cover type distributions can be computed as the database is queried.

- As new Bayesian networks are added to the system, they are linked to the data in the database, and are immediately available for use in the GUI application.
- A client application makes requests to the maps using the standard WMS protocol[5]. This means that information from other servers besides the CLIME-DSS server could, in principle, be accessed by the GUI application.

21.5 Examples of Using the CLIME-DSS

The following examples illustrate the use and capabilities of the CLIME-DSS. The water quality examples are based on hypothetical queries and should not be considered as real projections for the selected locations. The climate data and the BNs used are those described earlier in this chapter. Some of the queries require no site-specific inputs but most require further details of critical attributes that can either be actual values or supplied as 'guesstimates'.

21.5.1 Example 1: Using the CLIME-DSS To Show the Projeced Change in the Distribution of Winter Precipitation Across Europe

In this example we show how a user can explore the projected changes in the climate on a pan-European scale. In Fig. 21.11, the GUI in the DSS has been used to display the distribution of the relative change in winter precipitation together with more detailed information on the changes projected for Helsinki area. The user has clicked on the map to select Helsinki, which is then highlighted by an orange dot. The displayed results are from the RCAO-E climate model driven by the A2 emission scenario.

In the map, blue indicates a decrease and red an increase in the projected precipitation. The map shows a pronounced increase in the winter precipitation for Helsinki and the time-series graph shows that the seasonal variations are also very different with large increases suggested for the winter. The tables show the projected change in the other climatic variables. The average values of the variables are shown in the upper table and the proportion of the different types of day in the lower table. The tables allow the user to compare the current climate with the climates projected for two emission scenarios (A2 and B2).

By selecting different locations and switching between the RCMs, emission scenarios, and selecting either relative or absolute change the user can get a good impression of the changes projected for different parts of Europe.

[5]http://www.opengeospatial.org/standards/wms (accessed 23.2.2008)

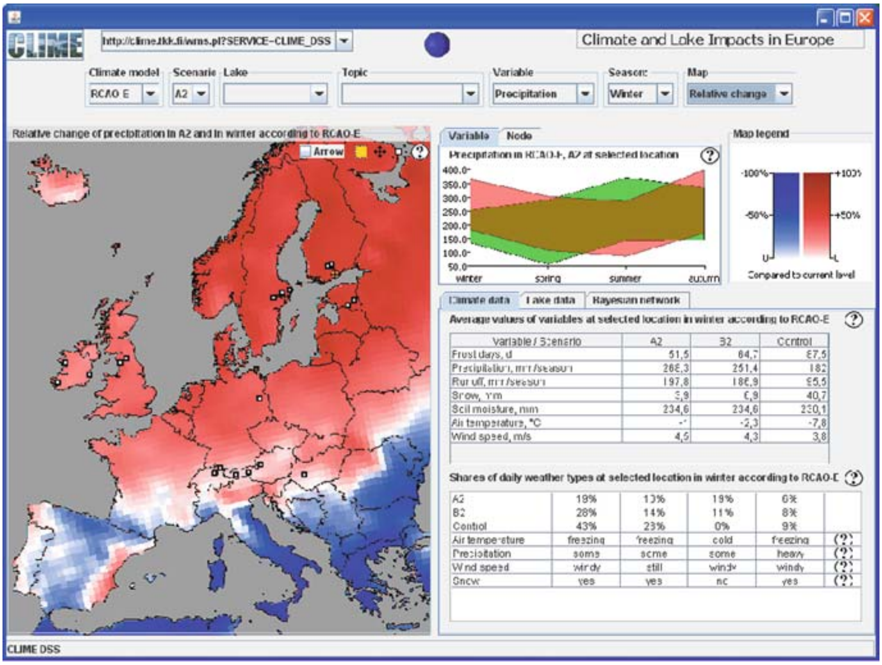

Fig. 21.11 Using the CLIME-DSS to explore the projected change in the distribution of mean winter precipitation in Europe and quantifying the specific changes projected for the Helsinki area

21.5.2 *Example 2: Relating the Climate Projected at a Selected Location with the Current Climate Experienced at Other Locations*

The GUI in the CLIME DSS can also be used to explore the general pattern of ‘climate migration’ in Europe by quantifying the similarity between the projected climate at one location with the current climate at another location.

Figure 21.12 shows the location of two grid squares where the current climate is very similar to the projected climate in the Copenhagen area under RCAO-E. The left map shows that, under the A2 scenario, the projections for Copenhagen are very close to the climate currently recorded in Brittany. The right hand map shows the corresponding comparison for the B2 scenario where the projected climate for Copenhagen resembles the current climate in the Brussels area. The degree of similarity between the current and future climates is shown by the relative size of the white and red disks. When the disk is completely red, the similarity is 100% so the exposed area of white is a measure of the mismatch between the two climates.

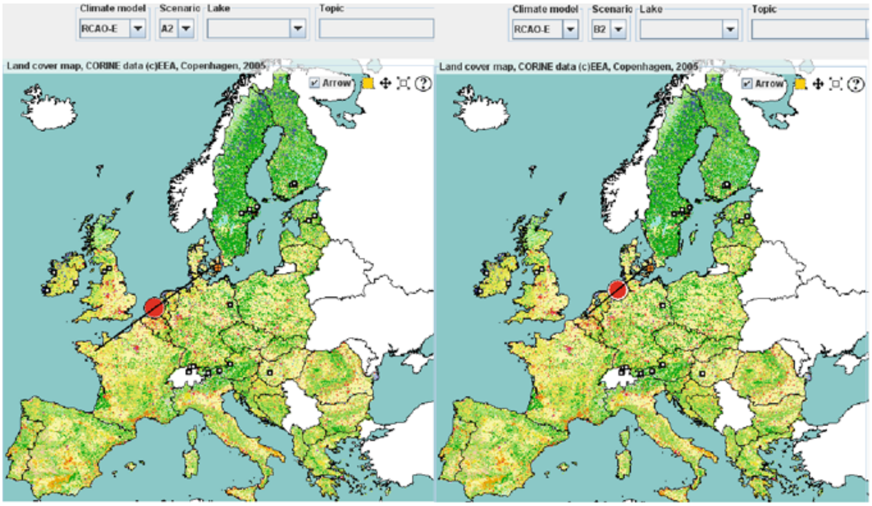

Fig. 21.12 The climate migration tool that forms part of the CLIME-DSS. The picture is a composite of two screenshots

21.5.3 Example 3: Using the CLIME-DSS To Quantify the Effect of Climate Change on the Timing of ICE-OFF

The CLIME-DSS can be used to predict how the projected changes in the climate might affect the timing of ice-off in European lakes. The prediction is based on the BN described in Section 21.3.3. In this example, we show how the DSS can be used to quantify the projected changes in the timing of ice-off in Pääjärvi, a large lake in southern Finland (Fig. 21.13).

The long-term records of the timing of ice-off from Pääjärvi can be used to set the 'baseline' conditions for the BN. Figure 21.13 shows that the user has selected *Ice cover* as a topic, chosen RCAO-E as the regional climate model and then selected emission scenario A2. When a topic is selected, the corresponding BN appears into the *Bayesian network* panel, and a *Calibration from* drop-down menu appears on the screen. The *Calibration from* menu allows the user to select the most appropriate network for the selected site. In this case, there is only one option in the drop-down menu (*Swedish lakes*) since these were the CLIME lakes with the most comprehensive 'ice-off' results. When the user selects a lake from the *Lake* menu, a red square shows its location and a BN diagram is displayed to show how the change in the timing of ice-off is related to the number of frost days at this location. The change (deviation) in the timing is then added to the probability distribution and the probability distributions for both the current and projected values appear in the *Node* panel.

The projections for Pääjärvi indicate that, in future, the ice will melt much earlier in the year. Existing records showed that in 8 out of 100 years the day of ice-off falls between Julian day 30 and Julian day 60, in 37 out of 100 years the limits are Julian

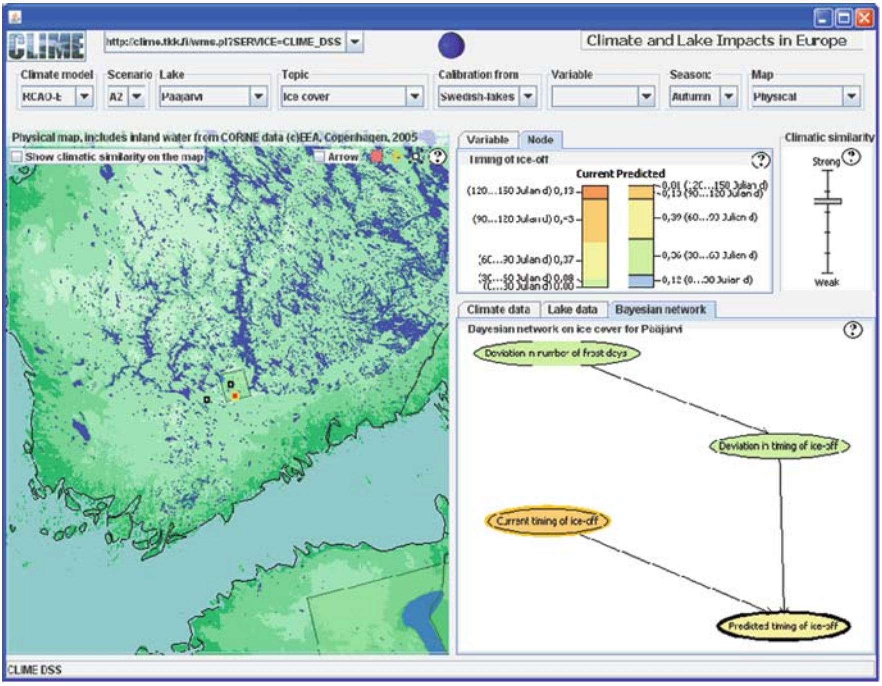

Fig. 21.13 Using the CLIME-DSS to predict the effect of climate change on the timing of ice-off in a Finnish lake. The *red square* shows the location of the selected lake

day 60 to Julian day 90 whilst in 13 out of 100 years it can be as late as May (Julian day 120 to Julian day 150). The corresponding numbers in the projection are 36, 39, and 1. In addition, the projection shows that in 12 out of 100 years the ice will melt in January (Julian day 0 to Julian day 30).

21.5.4 Example 4: Using the CLIME-DSS To Estimate the Mixing Depth of a Lake

Figure 21.14 shows the result of using the CLIME-DSS to quantify the effect of the projected variations in the climate on the summer mixing depth of Windermere in the English Lake District. The BN is described in Section 21.3.5 and includes values of the deviations in the air temperature and the wind speed. The nodes in the *Node* panel are the current and projected mixing depth. The climate scenarios suggest that summers in this region will become warmer and calmer but the amplitude of the seasonal variation in the wind speed is also likely to increase. The projected changes in the depth of the thermocline reflect this variability, and suggest that periodic reductions in the depth of the thermocline will be accompanied by periods of more intense mixing.

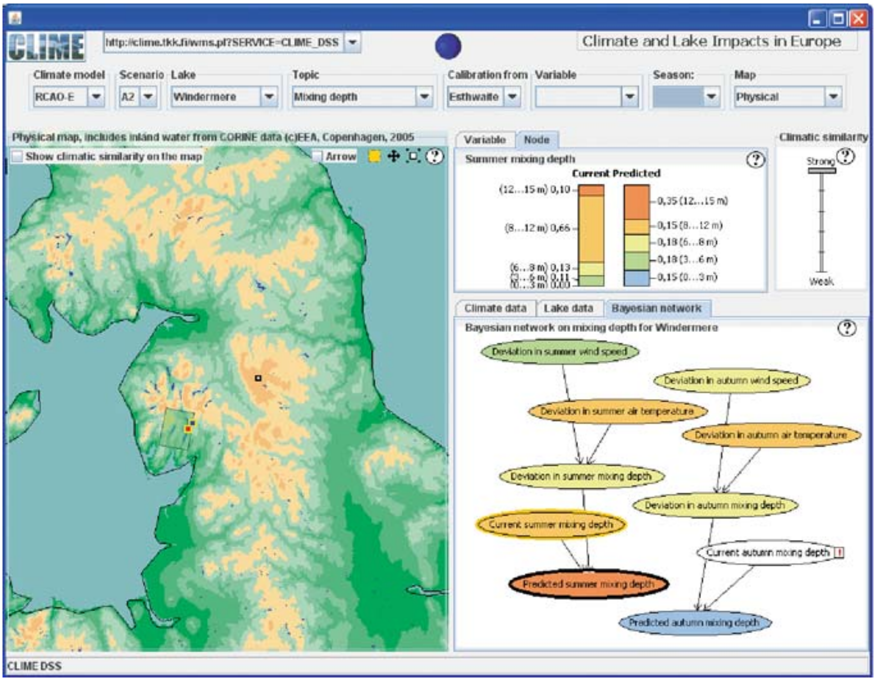

Fig. 21.14 Using the CLIME-DSS is to make projections for the average summer mixing depth of Lake Windermere

21.5.5 Example 5: Using the CLIME-DSS To Quantify the Effects of Climate Change on the Concentration of Doc

The CLIME-DSS contains a Bayesian network for exploring the potential effects of climate change on the flux of DOC from forested and peat catchments in Northern and Western Europe. In this example we show how this was used to quantify the potential effects of climate change on the flux of DOC from a peat catchment at Lough Leane, a lake situation in a mountainous area in Southwest Ireland. The BN for DOC is described in Section 21.3.8 and includes values for the deviation in the air temperature, soil moisture and run-off.

Figure 21.15 shows how *Water colour* was selected as the topic and Lough Feeagh as the best site for parameterizing the model for a site situated in Western Europe. Figure 21.16 also shows that the *Projected spring DOC* node has been selected in the BN before setting the current concentrations in the bar chart in the *Node* panel. These are centred on the 5–10 mg/l and generate a projected output that has a significantly higher mean and a less skewed distribution.

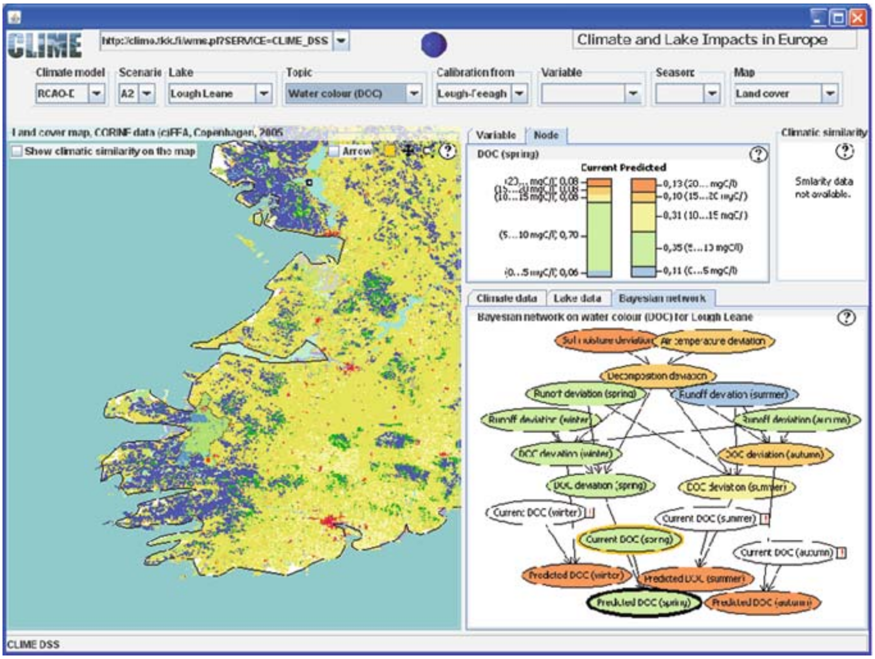

Fig. 21.15 Using the CLIME-DSS to explore the effects of climate change on the concentration of DOC in Lough Leane

21.5.6 Example 6: Using the CLIME-DSS To Assess the Risk of a Cyanobacteria Bloom in a Productive Lake

The BN for cyanobacteria blooms was described in Section 21.3.9 and includes values for the deviations in the summer air temperature, the run-off and the wind speed. Figure 21.16 shows the result of using this BN to assess the risks of a bloom in Lake Erken, a relatively productive lake 50 km north of Stockholm, Sweden. In this case, the only parameterization available was that for Esthwaite Water in the UK. The results for the RCAO-E model driven by the A2 scenario show a greatly increased risk with the summer concentrations of cyanobacteria increasing from a current range of 24–32 μg/l to a future range of 32–40 μg/l.

21.5.7 Example 7: Using the Network To Produce Projections for Sites Other than the Calibration Site

When the BNs in the CLIME-DSS are used to produce projections for sites not included in CLIME, we need to select an appropriate model and produce some measure of the confidence in the results. As a first approximation, we have assumed that

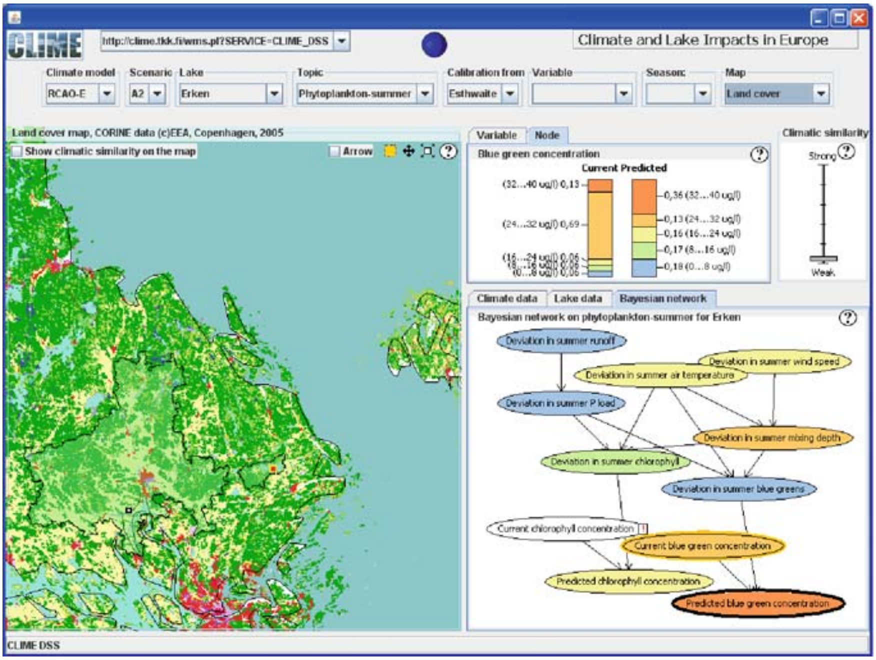

Fig. 21.16 Using the CLIME-DSS to quantify the effects of climate change on the risk of summer cyanobacteria blooms in lake Erken in Sweden (shown by the *orange square*)

same basic relationships will apply if the site of interest is located in an area where the climate is similar to that currently experienced at the calibration site. To help the end-user select the most appropriate calibration site, the CLIME-DSS includes a *Climatic Similarity* tool.

By checking the *Show climatic similarity on the map* checkbox in the upper left of the screen, the system displays a map that 'greys out' the areas in the map where the climatic similarity is deemed too low for a valid extrapolation. Figure 21.17 shows an example of such a 'confidence' map where the site of interest is Poulaphuca and Lough Feeagh has been selected as the calibration site. The map shows the geographic region, where current climate is rather similar to the current climate of South Western Ireland. The similarity is derived using the computation scheme described in Section 21.2 for the climate migration tool.

21.6 Conclusions

There are several issues that need to be addressed when developing a software system for assessing the impact of climate change on lakes. The CLIME-DSS is an information system designed to transfer scientific knowledge to the user in a flexible and intuitive way. The CLIME-DSS was produced using the state-of-the-art

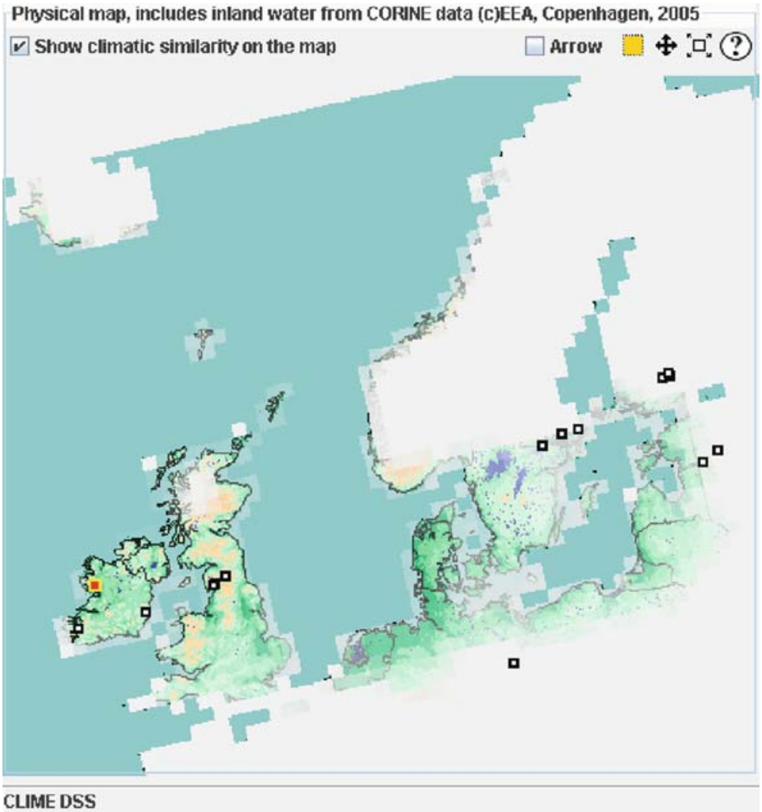

Fig. 21.17 A climatic similarity map generated by the CLIME-DSS. A *grey* mask covers the regions that are climatically very different from the Lough Feeagh area

technology and was deliberately based on existing high quality, free and open source tools.

Software and software systems require constant management to ensure their successful operation. In the case of CLIME-DSS, the system including its databases is still maintained by the original developers. The approach used in the CLIME-DSS is currently being adapted for use in other environmental applications. One recent development is its application in the EVAGULF project (Lindén, 2006), which applies BNs to an analysis of eutrophication problems in the Gulf of Finland. The flexible and extendable nature of web-based DSSs, such as the ones developed in CLIME and EVAGULF, allow new information and knowledge to be readily assimilated. Many of the tools in the CLIME-DSS are simple and use climatic variables that are available directly from the RCMs. The CLIME-DSS also allows the user to investigate the variability associated with different RCMs and explore the impact of these differences on the physiccal, chemical and biological characteristics of the lakes. The system demonstrates that the impact of the climate change is most visible in physical variables, such as the timing of ice-off, i.e., attributes that are directly driven by meteorological forcing.

The visualizations and data summaries available in the CLIME-DSS are already quite comprehensive. The coloring scheme in the maps has a limited resolution but the associated tables provide the user with more complete overview of the weather patterns projected for the different locations. The use of weather patterns (day types) for summarizing the projected changes differs from the more usual method of reporting seasonal averages but is more appropriate for situations where the attribute of interest is controlled by a combination of meteorological variables. The option of exploring the climatic similarities based on weather patterns could be of interest to researchers working in other fields.

At present, the scope of CLIME-DSS is confined to the comparison of the regional climates and their limnological impacts but the architecture could be extended to include the analyses of land-cover effects. The CORINE land-cover data is already included in the system as well as GIS functionality.

In CLIME, we used the outputs from a range of environmental models as well as the expert knowledge of the researchers to summarize the key findings as a series of thematically linked, probabilistic BNs. In the time available, we were only able to apply the simulation models to a limited number of sites but the structure of the CLIME-DSS allowed us to extrapolate these results to other locations in Europe. Since the sites covered by CLIME were all located in Northern, Western and Central Europe, the system also provides some guidance on where these extrapolations are considered appropriate. In principle, the methodology and technology that forms the core of the CLIME-DSS can easily be adapted to support other environmental projects, particularly those that involve large amounts of data and want to visualize the inputs and outputs in a geospatial domain.

In technical terms, the architecture of the CLIME-DSS has proved quite robust and relatively easy and flexible to manage. When the 'public' version of this system was implemented on the net, some users reported that the system was blocked by some firewalls. It is not very clear why this blocking occurs since the requests made by the Java client are similar to conventional web browser requests. The CLIME-DSS software is available for download at http://geoinformatics.tkk.fi/twiki/bin/view/Main/CLIMEDSS as a self extracting Microsoft Windows installation package. The web site includes a link to the source code of the software, which is also available under an open source license. This site includes the user manual, produced for an earlier version of the software, and notes on the system requirements for a successful installation.

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Chapter 22
The Socioeconomic Consequences of Climate Change for the Management of Water Resources

Ian J. Bateman and Stavros Georgiou

22.1 Introduction

In the CLIME project, a variety of analytical and modelling techniques were used to quantify the impact of climate change on lakes and reservoirs in Northern, Western and Central Europe. Chapters 3, 9, 11, 13 and 15 (this volume) describe some of the models used and Chapters 18, 19 and 20 the historical changes observed in the different regions. One issue, not addressed in these chapters, is the socio-economic consequences of the predicted changes in the quality of the water. Most of the lakes selected for intensive study were subject to a variety of human influences. Some, like Lake Erken in Sweden, are managed as water supplies. Others, like Lake Balaton in Hungary, are important tourist centres that contribute substantial amounts of revenue to the national budget (Várkuti et al., 2008).

From the outset, we identified two cross-cutting issues that were of interest to lake managers in all three regions. One was the increased frequency of algal blooms (Chapters 14 and 19 in this volume), the other was the progressive increase in the colour of water leached from upland and forest catchments (Chapters 12 and 13). In many lakes and reservoirs, the most serious water quality problems are not related to the inherent productivity of the sites but to the episodic appearance of algal blooms. At one time, such blooms were considered to be an inevitable consequence of cultural eutrophication but they are now known to be strongly influenced by year-to-year variations in the weather (George et al., 1990). The factors influencing the leaching of highly coloured water from upland and forest catchments are more complex. Mitchell and McDonald (1992) have shown experimentally that the drying and re-wetting of peat has a major effect on the amount of coloured dissolved organic carbon (DOC) released from the soil. The amount of colour

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Aquatic Ecology Series 4, DOI 10.1007/978-90-481-2945-4_22,

released appears to be related to the moisture content of the soil and the duration of the preceding drought. The most likely explanation for the increased production of coloured compounds is the aerobic decomposition of organic matter in the surface layers of the soil. The products of this breakdown are, however, only translated into colour on contact with water so the timing of release is strongly influenced by the hydrology of the catchment.

Socio-economic analysis can contribute towards improved management of these problems by informing decision-makers of the costs and benefits of mitigating the effects of climate change on the quality of water. In this chapter, different socio-economic valuation methods are used to assess the consequences of the expected increase in the frequency of algal blooms and the colour, taste and odour of the water delivered to consumers. The surveys and experiments organized to support these analyses were conducted in East Anglia, an area where the local population is familiar with algal blooms and where there are periodic complaints about the colour and taste of tap water. The 'water quality' projections used are based on historical observations at a number of CLIME sites, the model outputs and the climate change scenarios described in Chapter 2.

More detailed accounts of the socio-economic analyses described in this chapter have been given in a number of published papers (Bateman et al., 2004a,b, 2006a; Bateman and Georgiou, 2006). The aim of the present chapter is to summarize this information in a more accessible form and show how the various techniques can be used to quantify the effect of the projected changes in the climate on the management of water resources.

22.2 The Methods Used for the Socio-Economic Assessments

Natural scientists are still relatively unfamiliar with the techniques used to assess the socio-economic consequences of change in the environment. The essence of a socio-economic evaluation is to determine how society is affected by the assumed change in a particular environmental attribute. An essential first step is to describe this change in a way that can be readily understood by 'the man in the street' and then offer a range of options for managing the consequences. In this section, we describe some of the methods used in CLIME and show how these approaches have been used to assess the socio-economic consequences of changes in the climate on specific water resources.

Most of the lakes monitored and modelled in CLIME provide society with an important commodity (water) as well as a range of environmental goods and services. The projected changes in the characteristics of these lakes will have a direct effect on the quality of the water and an indirect effect on the systems capacity to support goods and services. The protection and maintenance of water quality almost always involves opportunity costs (foregone benefits). For example, there will be costs associated with limiting activities that have a deleterious effect on the quality of the water. In the socio-economic approach, we have to balance the opportunity costs of the resources used to protect water quality with the wider social (welfare) benefits provided by such protection. Such analyses serve not only to better protect water quality but also to improve the decision making process.

The economic basis of value is usually defined in terms of economic behaviour which is, in turn regulated by supply and demand (Freeman, 2003). Put simply, it is the maximum amount of a good, service or income that an individual is willing to forego in order to obtain some outcome that increases their welfare i.e., their willingness to pay (WTP). If the outcome reduces welfare, then this utility loss is measured by the minimum amount of money that the individual would require in compensation in order to tolerate the projected outcome i.e., the willingness to accept (WTA). Economic values are thus contingent on outcomes that are perceived as valuable to society. The outcomes themselves are therefore not necessarily of economic value. The value thus derives from the existence of a demand by society for the outcome e.g. the goods and services associated with good quality water. The total amount of money that society would be willing to pay or accept is the Total Economic Value (TEV) for the outcome being considered. The TEV is a composite measure that reflects all the different reasons and motivations that lie behind people's concept of value.

The economic value of the goods and services which would be lost or degraded as a result of climate change can then be contrasted with the associated costs (including the opportunity costs) of protecting/maintaining water quality within pre-defined limits. This approach is cost-benefit analysis (CBA) and is one of the methods used in any strategic socio-economic appraisal. In contrast to other methods, CBA is the only approach that compares costs and benefits in terms of a single common denominator (money). By adopting this approach, we can compare the economic efficiency implications of alternative actions (such as whether to protect/maintain water quality or not). In the case studies presented here, CBA was used to appraise a range of options designed to maintain/protect water quality from the effects of climate change. In this respect it entails the identification and economic valuation of all positive and negative effects of actions to protect/maintain water quality.

The first issue that needs to be addressed before these approaches can be applied to the kind of problems analyzed here is the valuation question. Many of the goods and services lost/degraded as a result of climate change are not priced in money terms. Indeed they have no market in which they are traded, and hence no market price to reflect their economic value. In such situations, a variety of techniques can be used to measure value in terms of willingness to pay (Fig. 22.1). These can be grouped into two basic approaches. 'Stated preference' methods that rely on data acquired by structured questionnaires where preferences are quantified via each individual's response to questions regarding hypothetical markets or choices. 'Revealed preference' approaches that infer values from individuals' market choices regarding goods which are related to the one being investigated, e.g. by looking at expenditure on holidays as a reflection of preferences for higher quality water.

Although a number of water quality issues have been addressed using the 'Revealed Preference' method (Bockstael and McConnell, 2007), the 'Stated Preference' techniques provide much greater flexibility in terms of the range of impacts that can be valued. As such, these techniques were chosen for this study in order to

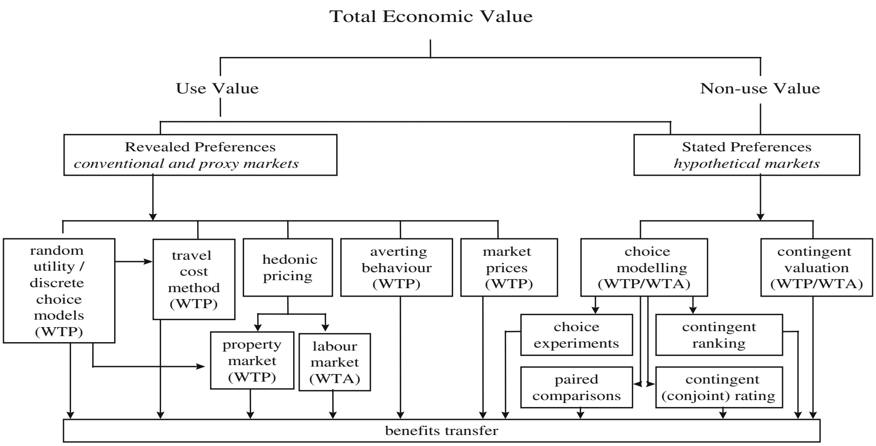

Fig. 22.1 Economic valuation techniques

assess the value of the goods and services which would be lost/degraded as a result of the impacts of climate change on water quality.

In practice, we used two different types of stated preference techniques: Contingent Valuation and Choice Modelling. The Contingent Valuation method (Mitchell and Carson, 1989; Bateman et al., 2002) uses a questionnaire survey to establish the individuals Willingness to Pay (WTP) for a single, specified change in the quality of water. So long as people are able to understand clearly the change being offered, and answer truthfully, this approach measures precisely what the analyst wants to know: i.e. the strength of the individual's preference for the proposed change. However, there are several practical difficulties with this approach, a central issue being whether the intentions indicated *ex ante* accurately describe the behaviour *ex post*, when people face no penalty or cost if their actual response is different to that indicated.

Choice modelling methods are most appropriate when a project or policy affects a particular aspect of a resource (Bennett and Blamey, 2001). Data collection is again via a survey during which respondents answer a series of questions, each being a choice between two or more options. By varying these options the analysts can see how respondents value the different characteristics (referred to as attributes) that define water quality. This permits valuation of marginal changes in those attributes (e.g., smell and taste, colour, etc).

The application of these two valuation techniques also requires the use of focus groups in order to identify, understand, and explore people's general knowledge about the subject and their perceptions of the issues being addressed. Focus groups are carefully planned discussions designed to establish people's perceptions on a defined area of interest in a permissive, non-threatening environment (Krueger, 1994). They are particularly effective in providing information about why people think or feel the way they do as well as an insight into why certain opinions are held. The key challenge in designing these experiments is to define how much information to present, how to present this information and how to refine the questions used in the subsequent valuation surveys. The aim is always to make the valuation scenarios as credible and realistic as possible.

Given this general discussion on the methodological approach, the focus of the case studies presented here is on the application of these techniques to value the potential impact of climate change on the water quality issues identified in the introduction.

22.3 Case Study 1: A socio-Economic Assessment of the Impact of Climate Change on the Increased Frequency of Algal Blooms

In the first case study, we assess the impact of climate change on the frequency and severity of algal blooms in lakes using as examples, photographs of a blue-green (cyanobacteria) bloom in the East Anglian area (see Bateman and Georgiou, 2006). We report the results of a cost-benefit analysis of a management strategy designed to mitigate the enrichment of surface waters, a factor that greatly increases the risk of such blooms. In order to carry out a cost-benefit analysis it is necessary to estimate the economic costs and benefits of preventing excess algal growths in lakes and reservoirs. Information on the costs of preventing nutrient enrichment is already available from secondary sources (see later). However, estimates of the economic benefits lost through the qualitative and quantitative effects associated with the periodic appearance of excessive growths of algae are not available from market price data. Consequently, the estimation of benefits forms the major empirical focus of this case study.

In order to obtain these values, a survey questionnaire based on the Contingent Valuation method was used to estimate individual's WTP for a scheme that would minimise the excessive growth of algae in a lake and ensure its continued use as an amenity/recreational site. In this instance, the respondents were asked to pay for a sewage treatment program that would remove phosphorous and reduce the frequency of 'problem' algal blooms.

In this valuation exercise, it was first necessary to select a set of target lakes before defining the population of individuals affected. Due to funding and other practical constraints, it was only possible to explore the perceived effects on a series of shallow lakes in the East Anglian region. These water bodies attract very large numbers of recreational users and have been the subject of intense scientific study for more than thirty years (Moss, 1996, 1998). These sites are much shallower than most of the CLIME lakes but they represent a type of water body that is known to be particularly sensitive to the projected changes in the climate. It is acknowledged that the results will thus provide a lower bound on the total costs and benefits for all the sites included in CLIME. Such analysis nevertheless provides important information for the management of the problem at the regional level.

In designing the survey questionnaire, a number of issues had to be addressed in a step-by-step way. Four focus group sessions of between six and seven people were organised and conducted with groups of East Anglian residents. The focus group protocol paid particular attention to participants understanding of issues such as: the concept of ‘open water’; the uses made of ‘open water’; the frequency of these activities; the characteristics of ‘good surface waters’; the concept of ‘algae’; the causes of algal growth; the problems caused by excess algae; what defines a ‘bad’ lake; how these ‘bad’ qualities are rated and how both ‘good’ and ‘bad’ open water is defined.

The findings from the focus groups were then used to develop a draft questionnaire for the contingent valuation survey. This comprised a number of sections that established the respondents’ views of: – their present use of these water bodies; – their understanding of the processes by which water bodies can be affected by algal growths; – their assessment of how such changes might impact upon the future usage of these water bodies. It also established their response to: – a valuation scenario that outlined the proposed remedial scheme; – a valuation exercise to determine households WTP to avoid the specified impacts; – belief indicators regarding perceived credibility of the proposed phosphorus reduction scheme.

Since the study was concerned with minimising the risks posed by the enrichment of lakes by treated sewage effluent, the valuation scenario paid particular attention to the combined effects of the climatic and demographic changes projected for the East Anglia region. Pictograms and photographs were used to explain the scenario (see Bateman and Georgiou, 2006) and revised as the focus group discussions progressed. Regarding the constructed market solution to the problem of perceived enrichment, survey respondents were asked to consider a plausible phosphate removal scheme at a local sewage treatment works. They were told that such treatment would increase their annual household water bill. Such bills have desirable properties as a payment vehicle since their compulsory nature minimises the risk of any ‘free riding’ behaviour.

After the presentation of the valuation scenario and payment vehicle, the elicitation question asked respondents how much they would pay to purchase the proposed scheme, under specified terms and conditions. The question was based on the application of a novel ‘one and one half bound’ (OOHB) dichotomous choice (DC) method for the elicitation of WTP values that was recently developed by Cooper et al. (2002). This presents the respondents with upper and lower bound limits of cost (bid) per household (or per individual) associated with provision of the good in question (see Fig. 22.2).

The elicitation questions were followed by a debriefing session where the respondents were asked to state their reasons for accepting or refusing the bid amounts. This was followed by questions about the respondent’s beliefs regarding the plausibility of the scenario presented. Finally, the survey concluded with questions concerning demographic and socio-economic characteristics of the respondent.

The draft questionnaire was employed to interview a pilot sample of around 100 respondents and the results used to specify the vector of bid values for the main survey. Results from the pilot survey also helped to ‘fine-tune’ the questionnaire. The main survey was carried out over a five week period using respondents from and around Norwich, the largest city in the East Anglia region. The survey was carried out using face-to-face interviews in accordance with the NOAA Panel

Phosphates can be removed at sewage works but this treatment costs money and would raise household water bills. We want you to consider a scheme which would pay for this treatment and avoid excess algae in the lakes and rivers of the region.

If this scheme was put in place, it would start working next year. While you would still pay your water bill in the same way you do now, these bills would go up.

Calculations based on engineering costs have established that the annual addition to a household's water bill would be between £ X and £ Y every year, starting at the beginning of next year.

Now I am going to ask you questions about whether you would pay either of these amounts but before I do want you to keep the following in mind:

Studies have shown that when people are asked about whether they would pay for environmental schemes, such as this one, they often say yes at the time they are surveyed, but later on wish they had said no when the time actually comes to pay. This is often for good reason, because people then realise that this will take money away from other things that are also important to them.

So, when considering what you would be prepared to pay for this scheme, I want you to think carefully about whether your household really would prefer to pay for this scheme, or would prefer to continue purchasing other things that are important to you. Remember that this money would be coming out your pocket and that would mean there would be less money for you to spend on other purchases that you might like to make.

1: **If the extra amount on your household water bill was £ X per year which of these options would you prefer?**

1 = Would pay (go to question 2)

2 = Would not pay (end of valuation questions)

Option 1: An addition to your water bill to pay for the phosphate removal scheme which prevents excess algae.

or

Option 2: No addition to your water bill so the phosphate removal scheme is not put in place and excess algae occurs.

2: (*Only ask to respondents who answered WOULD PAY to question 1.* **And if the scheme cost £ Y per year would you prefer to have the phosphate removal scheme or not?** (circle response number)

1 = Yes, Would pay (end of valuation questions)

2 = No, Would not pay (end of valuation questions)

Fig. 22.2 The valuation question used to elicit Willingness to Pay (WTP)

recommendation (Arrow et al., 1993). To ensure high quality interviews, a team of experienced and trained interviewers was used to conduct the interviews in the respondents' homes. Respondents' addresses were selected on a non-probability cluster basis, designed to guarantee a selection of households with different socio-economic and demographic characteristics.

A total of 2,321 households were approached for interview and 1,067 refused to participate in the survey, giving a response rate of 54%. A total of 1,254 households completed the survey questions, the results of which were used to calculate Willingness to Pay. This represents one of the largest surveys of its type in Europe.

The sample exhibited substantial variation across socio-demographic characteristics, permitting the analysis of a number of covarying influences upon WTP responses. Although the survey did not set out to capture a representative sample of the region, comparison with official statistics suggests that the sample was reasonably representative. With regards to the main part of the valuation exercise, that of obtaining estimates of the mean and median WTP values for the phosphorous removal scheme, parametric regression modelling of the OOHB responses was undertaken. Details of the modelling procedures are given in Bateman et al. (2004b).

The coefficients from this modelling procedure were then used to estimate the mean and median WTP per household for a scheme that minimised the risk of enriching the surface waters of East Anglia. The mean household annual WTP for the sample was found to be £75.41. In aggregating the benefit estimates for application within a Cost Benefit Analysis this WTP values is multiplied by the number of people that benefits from the reduction of enrichment, in this case the population of the East Anglian region. The results of the benefits aggregation exercise are shown in Table 22.1.

The estimates differ according to the precise WTP values assumed to apply. In this respect, two possible alternatives were identified. For the first estimate, no adjustment was made to the sample mean WTP to take account of those respondents who refused to be interviewed (i.e., the sample mean WTP was applied to the whole of the target population). In this case, the population aggregate WTP was estimated by multiplying the sample mean WTP (£75.41) by the number of households in East Anglia (2.253 million) to obtain annual benefits of £170 million. The second estimate assigned a zero WTP to those who refused to be interviewed (as per Bateman and Langford, 1997). As such, the inclusion of the 1,067 respondents who refused to be interviewed results in a lower sample mean household WTP per year of £38.48 giving an aggregate annual benefits estimate of £87 million.

Comparing the annual aggregate benefits of preventing the nutrient enrichment of lakes in the East Anglian region by implementing a phosphate removal scheme (Table 22.1) with the current annual costs of addressing nutrient enrichment damage for the whole of England and Wales (reported at £54.8 million – see Pretty et al., 2002), it is clear that net benefits are positive. Even for the lower bound benefit estimate value of £86.7 million/annum, the current costs of addressing enrichment

Table 22.1 Aggregate benefits of the phosphate removal scheme for East Anglia region

Aggregation approach	WTP per household (1) (£/year)	Aggregate Benefits (£) [(1) × 2.253 million]
Aggregation with sample mean WTP (excluding non-response) and East Anglian population – total sample	75.41	169.89 million
Aggregation with sample mean WTP (including non-response) and East Anglian population – total sample	38.48	86.70 million

for the whole of England and Wales are less than the perceived benefits of preventing enrichment just for the East Anglian region.

Although there are a number of uncertainties associated with both the cost and benefit estimates, fresh water resources would appear to be highly valued, and policies and schemes to protect them considered worthwhile investments. This point is driven home by the fact that the benefit valued in this study was only the prevention of algal blooms. Other benefits linked to the protection and management of rivers and lakes, such as the value of water bodies for commercial and other uses, strengthen the argument for investing in effective phosphate removal schemes even more compelling.

22.4 Case Study 2: A Socioeconomic Assessment of Climate Change Impacts on Domestic Water Supply Quality Changes

The second case study assesses the socio-economic impact of the projected increase in the leaching of dissolved organic compounds into waters abstracted for domestic consumption as summer droughts become more frequent. We report the results of an economic valuation study undertaken to assess the management options available for mitigating the effects of a sustained increase in water colour. In particular, the study considered the benefits that individuals in the East Anglia region would derive from preventing any increase in discolouration and any odour or tastes associated with the subsequent treatment of water.

In order to obtain these values, a survey questionnaire based on the Choice Experiment (CE) approach was used to estimate individual's WTP for a scheme that would prevent any sustained deterioration in the colour of water, i.e., an increase in the number of days when tap water had a 'rusty' or brown colour and/or smelled and tasted of chlorine. Once again, the valuation exercise was only concerned with an assessment of the problem for the East Anglian region.

Focus group discussions and pilot testing were again used to support the design of the CE questionnaire. The focus group sessions were very similar to those discussed earlier, but the protocol now focused on defining the characteristics of 'good' and 'bad' tap water,. In particular, participants discussed what are the indicators of 'good' tap water; a scale of 1 (worst) to 10 (best) for drinking water in the region; and the problems that could arise in both the medium (10 years) and longer term (30 years).

Based on previous work by the authors (Bateman et al, 2002, 2006b), a hypothetical choice experiment valuation scenario was developed, presented to participants in the focus groups and then used as the basis for a large-scale survey questionnaire. The questionnaire was composed of three parts. The first part was concerned with respondents' usage and perception of tap water. We asked respondents about the frequency of any tap water problems, their expenditure on bottled water and the proportion of water consumed from bottled sources, home-treated sources and directly from the tap. The second part contained the choice experiment and choice tasks.

Pictograms were again used to explain why and how water quality might change in the future (see Bateman and Georgiou, 2006). The last part of the questionnaire included questions related to the socio-economic characteristics of the respondents, such as gender, occupation and income.

Based on the information gained from the focus groups, the state of domestic water supply was described via three attributes. These were: (i) the number of days each year when a household's tap water smelled and tasted of chlorine; (ii) the number of days each year when the household's tap water was a 'rusty' or brown colour; and (iii) the additional annual costs incurred by the introduction of technical procedures to address these problems. Hose pipe bans were considered, but not used as an attribute since the pilot testing had shown that a very large number of days with hosepipe bans would be required to induce any trade-offs with the other attributes.

In the choice task, the respondents were asked to make a choice between a constant 'status quo' (SQ) and a varied 'alternative state' combination of attributes. The SQ was the level of tap water problems likely to be experienced in the coming year. In order to ascertain credible levels for status quo and the specified alternative state, initial attribute levels were piloted using a combination of e-mail surveys, face to face interviews and focus group sessions. This pilot testing indicated that the SQ levels could be set at 10 days per year for the smell/taste attribute and 5 days per year for the colour attribute (assuming no additions to the annual water bill). While pilot participants believed that technical solutions could be found to reduce or eliminate the smell/taste and colour problems, we did not investigate levels above the SQ since this might have given rise to credibility problems. As such, the 'status quo' therefore defines the upper levels of our smell/taste and colour attributes. These and other levels of the attributes (including non-zero increases in water bills to pay for reductions in smell/taste and colour levels) are shown in Table 22.2.

Table 22.2 Attributes and levels

Attributes	Levels
Number of days each year on which your tap water smelled and tasted of chlorine (ST)	0, 3, 6, 10 days
Number of days each year on which your tap water was a 'rusty' brown colour (C)	0, 1, 3, 5 days
Addition to your annual water bill	£10, £20, £30, £50

As mentioned above, a simple choice format was adopted in which subjects were faced with choosing between the SQ (held constant across all choice tasks), labelled as the 'No Scheme' state, and an alternative scenario described as 'Scheme A'. An example is shown in Fig. 22.3. Different scenarios were used in each successive task, these being labelled B, C, D, etc. Prior to the initial choice task, respondents were shown details of all the attributes and levels used in the study. Given that there were four levels of choice for each of our three attributes, the number of possible combinations is 4^3 i.e. 64 combinations. Each respondent was asked to answer multiple choice tasks but the total number was limited to 17 so as not to tire the

	'No Scheme'	'Scheme A'
Number of days each year on which your tap water smelled and tasted of chlorine	10	6
Number of days each year on which your tap water was a 'rusty' brown colour	5	5
Addition to your annual water bill	£0	£10
	Choose 'No Scheme'	Choose Scheme A
Which would you choose? (*tick one box only*)	☐	☐

Fig. 22.3 An example 'Choice task' from the water colour and odour questionnaires

respondent. The choice task questions included a 'repeated choice' task presented as the first and last questions submitted to the respondents.

Once these methodological design issues were resolved, the main survey was also conducted in the city of Norwich. Once again, a team of experienced and trained interviewers was used to undertake face to face interviews at the respondents' homes. Respondents' addresses were again selected on a non-probability cluster basis and sampling was undertaken over a nine week period during which time 864 questionnaire interviews were completed. The sample was again considered to be reasonably representative of the regions population.

In order to obtain estimates of the mean WTP values for the scheme designed to avoid the problems associated with discolouration, random utility modelling of the choice task responses was undertaken (McFadden, 1974). Further details of the modelling procedures are given in Bateman et al. (2006a). The estimates of households mean WTP for the avoidance of the tap water problems associated with increased climate change are as shown in Table 22.3.

Table 22.3 Estimates of WTP for the avoidance of tap water problems

Mean WTP (£) per household to avoid one day of colour problems	£5.40
Mean WTP (£) per household to avoid one day of smell and taste problems	£3.96

Unlike the first case study, no attempt was made to aggregate these per household estimates for the population to produce an overall Cost Benefit Analysis. This is because we do not yet know how climate change will impact on water colour in the Anglian region. Furthermore, in order to compare these benefit estimates within a cost benefit analysis we would also need more information on the costs associated with any remedial schemes. Once again, no specific estimates or predictions are available. Nevertheless, the benefit estimates obtained in this study indicate the high value people attach to water quality. These are likely to be substantial when multiplied by the population to provide aggregate benefits estimates, even if the improvements are relatively small. As such, any schemes to protect domestic tap water quality from the likely impacts of climate change are again considered worthwhile and are likely to represent a wise investment.

In addition to the two valuation case studies described above, a number of methodological and validity issues were also investigated (see Bateman and Georgiou, 2006 for details). These addressed some of the problems commonly associated with the design and analysis of the valuation studies, as well as some general issues connected with the use of monetary valuation techniques in making public policy decisions. Concerning the procedural aspects of the elicitation of individuals' valuation estimates, the results provide guidance to valuation study practitioners on the appropriate design and analysis of such studies, as well as their limitations in trying to obtain, robust and accurate estimates of economic value. For example, they consider the robustness of the 'one and one half bound' dichotomous choice elicitation method to anomalies (see Bateman et al.,2004b), and, how respondents' behaviour changes (to incorporate learning) over the course of repeated valuation choice exercises (see Bateman et al., 2004a). Inadequacies in the sampling methods currently used in contingent valuation studies were also investigated (see Bateman et al., 2006a). Once again the results are suggestive of more appropriate procedures for the production of policy relevant valuation estimates. Finally, there was an investigation of the preferences of different groups regarding the appropriate provision of environmental goods either through the market or public provision mechanisms. This issue goes to the heart of concerns about democratic government, and the nature and role of cost benefit analysis and monetary valuation techniques for making public policy decisions (see www.uea.ac.uk/env/cserge/enviro_valuation.htm).

22.5 Discussion and Conclusions

If present trends continue the direct and indirect effects of climate change will have a major effect on the management of freshwater resources which, in turn, will have financial, social and political impacts on all the sectors of the economy (see also Chapters 23 and 24 this volume). This management involves trade-offs between climate change impacts (such as the depletion and degradation effects considered in the case studies above) and the various uses that society makes of these water resources. Since economics is concerned with the concept of scarcity and with the mitigation of scarcity, it provides the essential framework for defining the conditions required to secure the most efficient use of these resources. In all cases, it is important to ensure that the 'true' economic value of a scarce resource is accounted for in the management of that resource. Unless these economic values are accounted for, there will be a misallocation of resources and sub-optimal societal wellbeing. Valuation of the benefits of freshwater ecosystems is particularly important for the following reasons:

1. These resources play a key role in national development strategies – depletion and degradation of lakes and freshwater resources imposes costs to nations, some of which produce impacts on GNP.

2. Mismanagement of these resources could impact on National Accounts – measures of economic activity which ignore the impact of climate change on water resources will fail to record an important element in the development of a sustainable economy.
3. More robust valuations are required to set national and sectoral priorities – as already mentioned, information on the economic value of policy changes can greatly assist government in setting policy and sectoral priorities.
4. The results can be used for project, programme and policy evaluation – the traditional role for environmental resource damage and benefit estimation is in project appraisal.
5. They form an important element of any proposed sustainable development programme – Once the goal of sustainable development is adopted, the requirement for valuation is somewhat greater than when considering efficiency alone.

The CLIME project was designed to provide strategic support for the Water Framework Directive, an area of legislation that is currently proving a major challenge to lake and catchment managers throughout Europe (see Chapters 1 and 23 this volume),. The analytical and modelling studies that formed the core of the project cannot, in themselves, be used to assess the sustainability of these systems from a societal point of view. In this respect, it is necessary to consider how socioeconomic systems interact and respond to changes. The socioeconomic perspective helps decision makers plan the efficient allocations of scarce resources so as to maximise the social wellbeing.

The case studies reported here explicitly consider the socioeconomic consequences of two water quality problems that are likely to become increasingly acute as the world becomes warmer. As such, they form an essential part of an overall impact assessment and show how selected ‘climate change’ outputs can be set in a socio-economically meaningful context. They also allow decision makers to consider the implications of these results for policymaking and management on a regional as well as local scale. Although research has recently been undertaken to support climate change policy at a macroeconomic level of analysis (Stern, 2006), such analysis has lacked the support of socio-economic impact evidence at the micro-level. The present chapter provides a first, albeit partial, attempt at the providing this evidence for two issues that have an important effect on our water resources.

The Case Studies also have important implications for the future development of water-related legislation in Europe. Given the importance of water to society, there are already statutory obligations and strong political pressure for the sustainable management of freshwater resources. These resources have been experiencing intense and sustained pressure from a range of direct and indirect driving forces. The stresses put upon the integrity of lakes and freshwater ecosystems are likely to be further exacerbated as a result of pressure from global climate change. The recently agreed EU Water Framework Directive (WFD) provides an important impetus for the sustainable management of lakes and freshwater ecosystems. Other important related EU policies include the Nitrates Directive (91/676/EEC), the Drinking Water Directive (98/83/EC), the Urban Wastewater Treatment Directive

(91/271/EEC), and the proposed new Groundwater Directive (COM(2003)550). Furthermore, reform of the Common Agricultural Policy, as well as the increasing pressure on public finances, are making it even more important to correctly allocate resources with respect to fresh water resource protection.

The outputs from the case studies presented here provide, for the first time, scientifically sound estimates of the socio-economic values associated with the degradation and use of freshwater resources. The first case study focused on the loss of aesthetic and recreation uses of lakes as a result of the projected increase in the frequency and severity of algal blooms (see Chapters 14, 15 and 19 this volume). The analysis of the costs and benefits of preventing these algal blooms demonstrated how even a 'crude' cost benefit analysis can play an important role in decision making. The suggested 'local' remedial measures were considered very worthwhile from an economic perspective and would have a major impact on societal wellbeing if similar investments were organized for the whole of the UK. In the second case study, the public attitude to the discolouration of water supplies provides water managers with an equally sound basis for future decisions. The ability to monetise the impacts of climate change in this way allows society to re-appraise the resources required to mitigate the effects of climate change but also to target these allocations at systems that are most likely to respond to the proposed management strategies.

The studies reported here also led to the development of new, improved methods for obtaining valid, reliable and accurate estimates of the economic values associated with the defined aquatic resources. These results should also help others design appropriately robust estimates of the economic value of natural resources.

From a macro-economic perspective, the case studies demonstrate that the impacts of climate change on freshwater ecosystems can impose significant losses on societal welfare that can only be avoided if appropriate counter measures are implemented in good time. Whether such measures are in the form of regional policies to mitigate specific impacts (e.g. WFD policy to tackle eutrophication) or global policies to prevent climate change (e.g. such as a carbon tax) will depend on the relative benefits and costs of the different policies. The analyses described here have obvious limitations since they focus on two, quite specific, climate-related problems. They are, however the first to assess the perceived value of the management options available to mitigate the specified impacts and could be extended to produce a more comprehensive assessment of the strategies used to combat the impacts of climate change on the aquatic environment.

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Chapter 23
Climate Change and the Water Framework Directive

Tom Frisk and Glen George

23.1 Introduction

The Water Framework Directive (European Commission, 2000) is a package of environmental legislation that establishes new principles for action in the field of water policy. The central aim of the Water Framework Directive (WFD) is to restore or enhance all aquatic habitats in the territory of the European Union by 2015. Surface waters are defined as having good ecological status if their physical, chemical and biological characteristics match those of a comparable water body subject to minimal anthropogenic influences. Member States are required to develop River Basin Management Plans (RBMP) to meet these objectives and update these plans at regular intervals.

In the last five years, a number of working groups have been established to provide technical advice on the implementation of the Directive. A number of Guidance Documents have also been produced which describe these principles in general terms. These include guidelines on the establishment of reference conditions and recommendations on the analysis of pressures and impacts (IMPRESS, 2003). All the guidance documents published to date are available on the internet and can be accessed via: http://ec.europa.eu/environment/water/index_en.htm. During this development period, the Commission also funded a number of scientific projects on topics connected with the implementation of the Directive. The CLIME project formed part of a group of projects called CatchMod (Blind et al., 2005) where the focus was on the development of tools and models for catchment management. Some of these projects addressed generic issues, such as the development of IT protocols for linking simulation models (www.openmi.org). Others dealt with more specialised topics, such as the ecology of rivers in arid zones (www.tempqsim.net) and the management of catchments in trans-boundary areas (www.tiszariver.com).

The primary objective of CLIME was to develop and test models that could be used to simulate the responses of lakes to future as well as past changes in the

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climate. The secondary objective was to describe and analyse the long-term changes recorded in a number of European lakes and relate these to regional and global-scale variations in the climate. At present, the WFD makes no explicit reference to the effects of climate change but this issue has been addressed in a number of reports and papers (Arnell, 2001; Eisenreich et al., 2005; European Environment Agency, 2007; Wilby et al., 2006). In this chapter, we highlight those aspects of CLIME that are most relevant to the implementation and development of the WFD. The project has already published more than a hundred papers on the potential effects of climate change. The Decision Support System (DSS) described in Chapter 21 is also available on line at http://geoinformatics.tkk.fi/twiki/bin/view/Main/CLIMEDSS.

23.2 Addressing the Uncertainties Associated with the Changing Climate

In Chapter 1 of this volume, George used a schematic diagram to explain how the water quality targets set by the WFD might have to change to take account of the new boundary conditions imposed by the changing climate. For water managers, the key task is to distinguish those changes that are primarily driven by local changes in the catchment from those that are driven by regional changes in the climate. It is now becoming clear that the methods used to support the WFD will need to be revised at regular intervals to accommodate both the direct and indirect effects of climate change. In CLIME, we identified six climate-related issues that merit consideration:

1. The methods used to extrapolate the results of climate models to a catchment scale.
2. The impact of the changing climate on the WFD typology and classification of lakes.
3. The value of long-term monitoring in climate impact assessments.
4. The modelling techniques used to assess the climatic sensitivity of lakes.
5. The conceptual model used to support the WFD.
6. The methods used to disseminate the results of climate-related research to a wider audience.

23.2.1 Downscaling the Results from Climate Modelling

Three-dimensional models of the earth's atmosphere provide the best physically-based means of describing the effects of the global increase in CO_2. In recent years, the resolution of these models has increased and more attention is now paid to the uncertainties associated with the different projections (Deque et al., 2005).

CLIME was one of the first projects to take advantage of the increased spatial resolution provided by the new generation of Regional Climate Models (RCMs). The RCMs used were the Rossby Centre Regional Climate Model (RCAO), developed by the Swedish Meteorological and Hydrological Institute, and the Hadley Centre Regional Climate Model (HadRM3P) developed in the UK. The parent General Circulation Models (GCMs) were the HadAM3H/P from the Hadley Centre and ECHAM4/OPYC3 from the Max Planck Institute in Germany. A general description of these models and the methods used to extrapolate the results to a catchment scale has been given by Samuelsson in Chapter 2. All those involved with applying these results in their impact assessments must take account of the uncertainties that are an inherent part of the modelling process. The first level of uncertainty is that associated with the choice of model. In CLIME, we used two GCMs, two RCMs and two different greenhouse gas emission scenarios. The second source of uncertainty is the stochastic nature of the weather and the effect that small differences in the initial conditions can have on the projected variations. These sources of variability are particularly important for impact assessments on lakes which are often very sensitive to short-term changes in the weather (see Chapter 16, this volume).

The weather generator software developed in CLIME proved an ideal means of quantifying this uncertainty by providing the modellers with the multiple sequences of daily weather needed to produce the stochastic outputs. The schematic in Fig. 23.1 shows the scheme that was usually used to drive all the models developed in CLIME. In a typical application, at least a hundred 30-year sequences of the daily weather were used to perturb the models and the procedure then repeated for all possible combinations of RCMs and emission scenarios. Stochastic models of this kind provide the key to quantifying the 'cascade of uncertainty' associated with the projected changes in the climate. In future, water managers in Europe will have access to high-resolution transient simulations of the changing climate i.e. scenarios

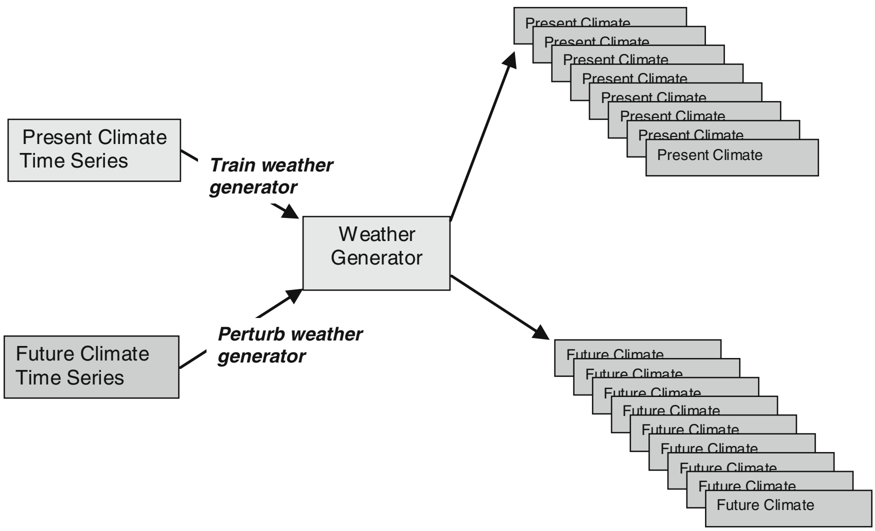

Fig. 23.1 Schematic showing the way the weather generator software was used to produce multiple sequences of the daily weather

that simulate the time course of change (Chapter 2, this volume). New techniques will then have to be devised to assimilate this large volume of data and assess its significance for the management of lakes.

23.2.2 The Impact of the Changing Climate on the Classification of Lakes

The basic 'building blocks' of the WFD are the River Basin Management Plans (RBMPs). In these plans, Member States are required to classify all the surface waters within their territory and assess their ecological status. These classifications are based on a series of hydromorphological, chemical and biological attributes that describe both the structure and the seasonal dynamics of the individual sites. Table 23.1 lists the key measurements used to classify lakes and assess their compliance with the WFD. Many of these variables will be directly influenced by changes in the climate whilst others will only be affected in an indirect way. The sites included in CLIME covered a range of lake types in eco-regions that extended from the west of Ireland to northern Finland. Significant climate-related effects were

Table 23.1 The water quality attributes used to classify the lakes and assess their ecological status. Those shown in bold were the ones selected for intensive study in CLIME

Biological elements		
	Composition, abundance and biomass of phytoplankton	
	Composition and abundance of other aquatic flora	
	Composition and abundance of benthic invertebrate fauna	
	Composition, abundance and age structure of fish fauna	
Hydromorphological elements		
	Hydrological regime	
		Quantity and dynamics of water flow
		Residence time
		Connection to the groundwater body
	Morphological conditions	
		Lake depth variation
		Quantity, structure and substrate of the lake bed
		Structure of the lake shore
Chemical and physico-chemical elements		
	General	
		Transparency
		Thermal conditions
		Oxygenation conditions
		Salinity
		Acidification status
		Nutrient conditions
	Specific pollutants	
		Pollution by all priority substances identified as being discharged into the body of water
		Pollution by other substances identified as being discharged in significant quantities

detected in the three CLIME regions with the most pronounced responses recorded in lakes that were already impacted by a variety of anthropogenic pressures.

One issue of critical importance to the Directive is the way in which these measurements are used to define the ecological status of the lakes. The Directive describes, in general terms, the characteristics of waters with high, good and moderate ecological quality. A key step in this process is the method used to assess the condition of the individual lakes. This is a ratio-based procedure where the ecological quality of each site is compared with that of a similar site that is assumed to be in a near-pristine state. The ratio is expressed as a numerical value between zero and one, with values close to one indicating a site with a high ecological status. Protocols for defining these reference conditions and setting the boundaries between high, good and moderate status have been developed and tested on a range of European lakes (REFCOND, 2003). Most of these protocols can be adapted to meet the challenge posed by the changing climate but only if enough information is available to define the new boundary conditions. This procedure for defining the ecological status of lakes on a pan-European scale has, however, proved controversial. Moss (2008) has argued that too much attention is being paid to the classification of 'types' at the expense of more realistic 'dynamic' concepts. Climate change, on the scale now envisaged, will clearly have a significant effect on both the system of classification and the way in which the limits are defined for the different lake types. The results of the analytical and modelling studies conducted by CLIME suggest that this procedure will now have to be modified to take account of three, climate-related, issues:

- The extent to which the reference conditions defined for each system will have to be modified to take account of the changing climate.
- The extent to which the ecological class boundaries may need to change to reflect a fundamental shift in the climatic boundary conditions.
- The extent to which the typology used to describe the lakes will have to be revised to account for the more extreme, climate-driven, variations in the status of the sites.

Owen et al. (2002) have argued that more flexible standards will have to be set before the present system can accommodate the pressures posed by the changing climate. More recently, Nõges et al. (2007) have suggested that the reference conditions defined for some sensitive sites now need to be reviewed and modified on a case-by-case basis. If the imposed climatic pressure has the same effect on the target and reference sites the ratio used to quantify its ecological status could remain the same. The dynamics of the two systems may now be very different but this shift will not be reflected in the statistic used to measure their ecological status. In most systems, the responses to the changed climate are likely to be non-linear and may cross thresholds that can be characterized as a change of regime. Regime changes of this kind are very often associated with a sudden change in the weather patterns experienced in a particular region. In the atmosphere, there is a tendency for the circulation pattern to operate in a small number of relatively stable configurations

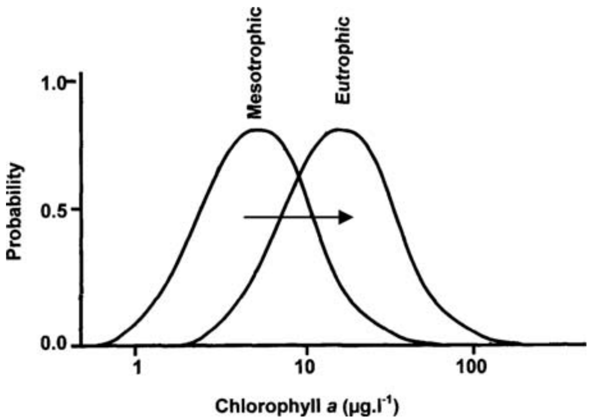

Fig. 23.2 The potential effect of climate change on the perceived trophic status of a lake

(Corti et al., 1999). Changes in the frequency of these circulation patterns, such as those described by George et al. in Chapter 16 and Dokulil et al. in Chapter 20, can either mask or amplify the effects associated with a gradual change in the climate. The schematic in Fig. 23.2 shows how regime changes of this kind could confound the classifications established to support the WFD. In this example, a lake originally classified as mesotrophic is shown to have acquired the characteristics of a eutrophic system due to a weather-driven change in its seasonal dynamics (see the examples presented by Nõges et al. in Chapter 14 and George et al. in Chapter 19). In CLIME, we also encountered a small number of cases where the projected changes in the climate would not only influence the classification of a lake but also its basic typology. The examples included a sustained change in the mixing characteristics of a Swedish lake (Chapters 8 and 15) and the projected increase in the colour of a lake situated in a peat catchment in southern Ireland (Chapter 13). In the guidance documents issued by REFCOND, it is generally assumed that the 'natural variability' recorded at a particular site will be much smaller than that of the parent 'type'. For some sensitive sites, this may not be the case in a warmer world e.g. if there is strong link between a water quality attribute and the frequency of an extreme climatic event. The Directive allows some quality elements to be excluded from the assessment if they display a high-degree of 'natural variability'. It is not yet clear if the extremes produced by the changing climate are covered by this technical proviso.

23.2.3 The Role of Long-Term Monitoring in Climate Research

In CLIME, we were able to take advantage of unique long-term records from lakes that had been the subject of intensive study for at least 20 years. The value of long-term data for climate-related studies cannot be overemphasised. Records of this kind provide a powerful way of assessing the inherent variability of the lakes and quantifying their response to extreme variations in the weather. A number of new approaches to the analysis of these patterns were developed during the course of this project. These included the 'weather typing' approach described in Chapter 16 and the Gaussian modelling approach described in Chapter 14. The time-series analysed in CLIME showed that lakes located in different parts of Europe responded in different ways to climatic forcing. In Northern Europe, the most important effects were those associated with the change in the freeze-thaw dates of the lakes and the resulting seasonal shift in their hydrological cycles (Chapter 18). In Britain and Ireland, the lakes were more sensitive to the flushing effects of heavy rain and the stabilizing effects of calm, anticyclonic weather (Chapter 19). The responses observed in Central Europe (Chapter 20) were more diverse since this region included some shallow as well as a number of very deep lakes.

One of the most challenging issues addressed by CLIME was the attribution question i.e. the extent to which the observed variations were related to local changes in the catchment or regional changes in the weather. When assessing the effect of external pressures on the status of lakes it is important to consider the ways in which they interact with other pressures on the system. The Venn diagram in Fig. 23.3 shows one way of summarizing the sensitivity of a lake to a combination of external pressures. The circles represent the potential effects of atmospheric deposition, catchment loading and the local weather on the ecological status of the lake. In CLIME, we used a combination of historical analyses and model simulations to quantify these pressures and identify the factors that were most likely to influence the ecological status of the lakes. One issue not addressed, was the effect that a systematic change in the climate would have on the management of land in the catchment and the types of crops grown. The ecological effects of the projected changes in the climate are much easier to assess than those connected with the management of the land. Changes in the climate may determine what crops can

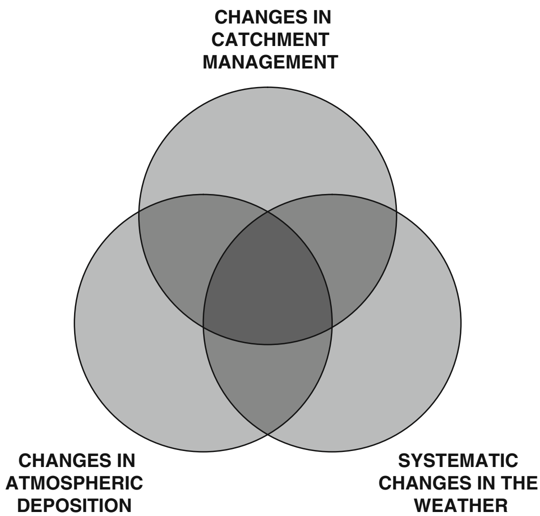

Fig. 23.3 Venn diagram showing the overlapping influences of changes in atmospheric deposition, catchment management and climate

be grown but what crops are actually grown will depend on a number of social, economic and political factors. For example, in Finland, where the growing season is short, agriculture is not very profitable. This situation could well change when the world becomes warmer (Hildén et al., 2005; Perrels et al., 2005) but the precise direction of change will be governed by economic and political factors e.g., the cost of water for irrigation and the support provided by government institutions.

23.2.4 The Conceptual Model Used to Support the WFD

One of the most important steps in the implementation of the Directive is the analysis of pressures and impacts. For most Member States, an inventory of pressures was completed in 2004 and guidelines then circulated to support their implementation. These recommendations were based on the familiar Driver, Pressure, State, Impact and Response (DPSIR) model. In this model, the 'Driver' is the anthropogenic change responsible for a particular effect, the 'Pressure' is the effect, the 'State' is the resulting environmental condition and the 'Impact' the ecological consequence. The 'Response' refers to the measures taken to counter the defined pressure and to return the water body to its reference state. In practice, it is much easier to measure the state than it is to define the pressure or quantify the impacts. The WFD assumes that the ecological status of a water body can be defined in relation to pressures that can be remedied by a local Programme of Measures (POM). When these pressures are associated with changes that operate on a global scale, remedial measures implemented at a local level may be so ineffective that the basic principles of the assessment are undermined. It remains to be seen whether new standards will be defined to accommodate these pressures or whether cases for a derogation (waiver) of the 'good status' objective will be advanced on a site-by-site basis. In practical terms, it is very important to target remedial measures at the most resilient sites, particularly where the cost of the remedial measures is relatively high i.e. the proposed actions must be affordable, proportionate and sustainable (Wilby, 2004).

In a warmer world, the pressures imposed by the changing climate could well eclipse those associated with the management of the catchment. The schematic in Fig. 23.4 shows one way in which a sustained change in the weather could compromise the measures taken to reduce the incidence of algal blooms. In this example, the solid arrows show the DPSIR sequence for the catchment and the broken arrows the regional effect of the changing climate. The key point is that the severity of the blooms is influenced by the combined effect of the local increase in the supply of nutrients and the regional effect of a systematic change in the weather. At one time, blooms of this kind were considered an inevitable consequence of nutrient enrichment but they are now known to be strongly influenced by variations in the physical characteristics of the lakes (George et al., 1990; Huisman et al., 2004). Bloom-forming species, like the cyanobacteria, grow quite slowly

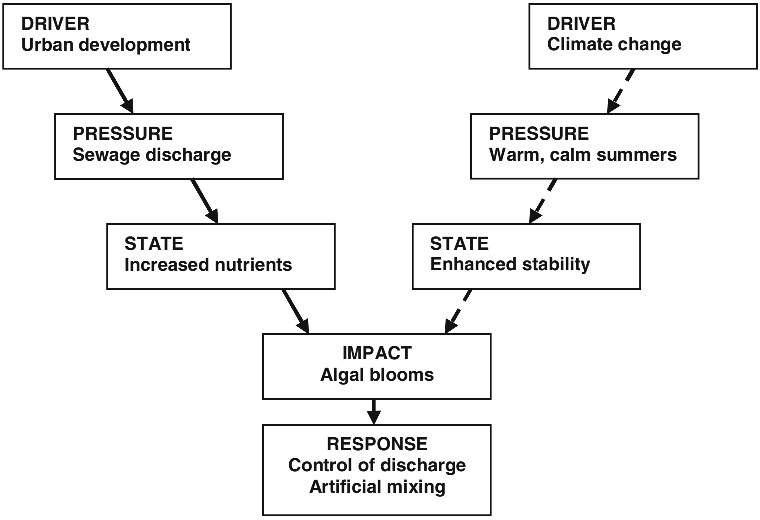

Fig. 23.4 Schematic diagram showing the combined effects of a local and regional driver on a DPSIR analysis of water quality

and typically only become dominant towards the end of the growing season. Mild winters and warm, calm summers accelerate this successional process and give rise to blooms that appear earlier and remain dominant throughout the summer. In some cases, it may be possible to reduce the severity of such blooms by introducing additional controls on the flux of nutrients. In others, special measures, such as artificial mixing, may have to be adopted to reduce the severity of the developing bloom (see Chapter 19 for a practical example).

23.2.5 Using Models to Assess the Climatic Sensitivity of Lakes

The Common Implementation Strategies for the WFD suggest that numerical modelling could play an important role in the assessment process (Frisk et al., 1994, 1997). The procedures currently used to support the WFD place too much emphasis on monitoring the structure of the systems rather than the processes that regulate their chemical and biological characteristics (Wilby et al., 2006). The most effective way of quantifying the climatic sensitivity of a lake is to model the processes that regulate the supply of nutrients. One problem with the model-based approach is the inherent complexity of many formulations and the detailed measurements required for their calibration. In general, the more complex the model, the greater the data requirements and the time and cost required for their effective implementation. In CLIME, we adopted what Carpenter (2003) called the ‘fast and frugal’ approach to catchment modelling. The models used were based on the

Generalized Watershed Loading Function (Haith et al., 1992; Schneiderman et al., 2002; Chapter 3, this volume). This uses a lumped-parameter scheme to simulate the critical transfer processes. Variants of this model were then used to simulate the flux of phosphorus (Chapter 9), nitrate (Chapter 11) and the supply of dissolved organic carbon (Chapter 13). In most EU-funded projects, such modelling tasks are sub-contracted to one or two expert groups. In CLIME, we adopted a devolved approach, where the modellers trained colleagues in the three regions to run their own simulations. In this way, many functional aspects of the models were improved and modelling skills acquired by new research groups in the Member States.

The other advantage of selecting such an easy-to-use model was the ease with which the modellers were able to run simulations with multiple realizations of the local weather. In some cases (e.g. DOC), this also allowed them to characterize the response of the systems to a number of extreme, but very rare, weather events. The confidence bands associated with these results also provided a measure of the uncertainty associated with these simulations and the climatic variations projected for the different regions. The examples in Fig. 23.5, show the results of using multiple simulations with GWLF to quantify the effect that the projected changes in the climate might have on the supply of phosphate to a Finnish and a British lake. The results suggest that the most important effects in the Finnish catchment will be those connected with a shift in the timing of nutrient loading. In contrast, the results for the UK suggest that the main difference will be the increased loadings of phosphorus projected for the autumn and winter. The recent increase in the number of climate scenarios available provides water quality modellers with the data they need to quantify the uncertainties associated with the projected change in the climate. The models selected must, however, be easy to use at sites that have not been subject to very intensive study.

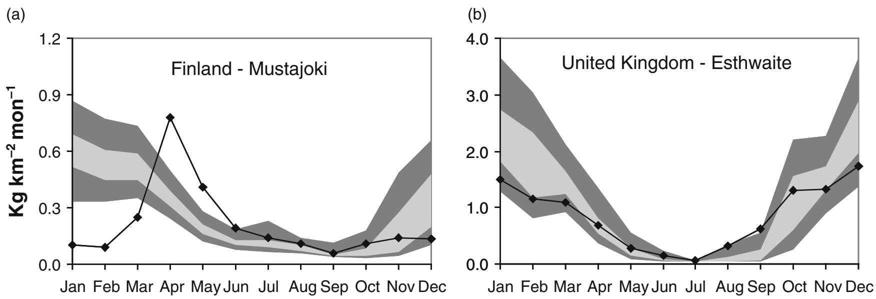

Fig. 23.5 A comparison of the observed and projected variations in the flux of phosphate from catchments situated in (**a**). Southern Finland. (**b**). Northern UK. The *line* shows the median of the control (present day) simulation, the *light grey* shading the range of the medians and the *dark grey* shading the range between the 25 percentile and 75 percentile limits in the six future climate simulations (see Chapter 9, this volume).

23.2.6 The Methods Used to Disseminate the Results of Climate-Related Research

The central concept in the WFD is the concept of integration. Water quality problems have, traditionally, been addressed on a case-by-case, issue-by-issue basis. The WFD provides a coherent, systems-based, structure for quantifying the status of lakes and combines these with recommendations on the practical measures required to sustain and improve water quality. A key objective of CLIME was to provide tools that could be used to manage lakes under future, as well as current, climatic conditions. The holistic approach adopted by the WFD makes it an ideal platform for addressing this issue on a pan-European scale. The monitoring programmes already in place (Irvine, 2004) provide the basis for this assessment but need to be combined with the predictions produced by the modelling tools developed to support the WFD (www.rbm-toolbox.net/bmw). Since such results are inherently complex, they need to be presented to the catchment managers in a readily accessible form.

One of the most important 'deliverable' in CLIME was the decision support system used to integrate the results collated from the different Workpackages. The schematic in Fig. 23.6 shows how these results were first assimilated, and then structured to form the basis of the CLIME Decision Support System (CLIME-DSS). The CLIME-DSS is a web-based application that allows a non-expert end-user to visualize the projected changes in the climate and assess their likely effect on a number of water quality variables. The Bayesian system that forms the core of the CLIME-DSS can also be used to assimilate and order information produced by other research programmes. A detailed description of this system is given in Chapter 21 of this volume. The latest version of the CLIME-DSS (http://geoinformatics.tkk.fi/bin/view/Main) allows the user to:

- Visualise the patterns of change projected for the European area using two climate models and two emission scenarios.
- Compare the climate projections stored at each map location to the current climate experienced at a warmer location.
- Visualise the potential effect of the projected changes in the climate on limnological variables such as ice cover, surface temperature, the supply of nutrients and the seasonal development of phytoplankton.

In addition, more expert users can interrogate the Bayesian networks that form the basis of the DSS to test a range of 'what if' hypotheses for their lakes and catchments.

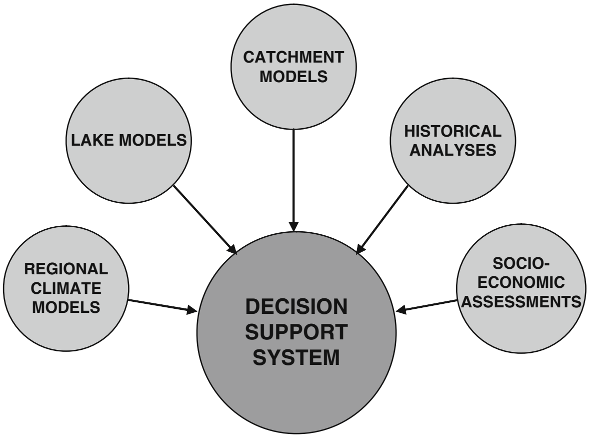

Fig. 23.6 Schematic showing how the results from the different CLIME Workpackages were assimilated by the Decision Support System (CLIME-DSS)

23.3 Discussion

Climate change is not explicitly included in the text of the Water Framework Directive. The WFD does, however, include procedures that can be used to address this additional pressure, notably the six-yearly reviews of the RBMPs. There is now no doubt that climate change, on the scale now projected, will have a major effect on the development of these plans. The current focus of the WFD is the control of point and diffuse source pollutants but pressures that fall outside these immediate aims will have to be identified, quantified and addressed. Discussions are currently underway on how best to meet the challenge of managing water resources in a warmer world. A Strategic Steering Group on Climate Change and Water was established by the EU and an overview of the key impacts prepared by the European Environment Agency (EEA, 2007). Successful adaptation to the changing climate will depend on the extent to which these pressures can be managed by a step-by-step modification of existing water regulations. The anticipated effect of global warming will affect most of the water quality attributes listed in the Directive and have a disproportionate effect on water bodies that are very near their type boundaries (Irvine, 2004; Nõges et al., 2007).

One question that has yet to be addressed is the extent to which the reference conditions used to define the ecological status of lakes will need to change to accommodate these pressures. If the reference and target systems respond is a similar way to the change in the climate, the metrics used as a measure of this status could remain the same. Such a ‘box ticking’ approach might meet the letter of the law but would not be a true measure of the functional response of the lake. To date, there is no indication that the pressures from climate change have had a significant effect on the achievement of ‘good ecological status’. This situation is bound to change by the third planning cycle, which is very close to the 2020 horizon used in many climate scenarios. In the intervening period, the Programmes of Measures already adopted will have to be revised to ‘climate proof’ the most vulnerable systems. In some cases, it may prove impossible to counter the effects associated with the progressive change in the climate. In such situations, it would be better to re-allocate resources to remedial work at less susceptible sites (see the ‘moving target’ analogy used by George in Chapter 1). At less severely impacted sites, the response will be governed by the cost of the remedial measures and the ‘willingness to pay’, a factor examined in Chapter 22. Pressures could well grow for a change in the review procedures, or even the underlying legislation. Changes of this magnitude clearly need to be evidence-driven and be based on the most up-to-date observations, models and climate projections.

In CLIME, we have made a start by analysing long-term data sets from a wide range of lakes and then perturbing our models with the outputs from several climate change scenarios. The geographic range of these analyses is, however, limited and there is an urgent need to expand these studies to surface waters in the Mediterranean Region. Here, the challenge is even greater, since most of the standing waters are reservoirs that are already subject to periodic droughts. Moss (2008) has described reservoirs as ‘disabled waters’ and lists a number of features, such as the extreme variations in the water level, that make it difficult to maintain any semblance of a ‘natural’ ecology. The standards set by the WFD for such heavily modified waters are less demanding but even these would be very difficult to sustain in a warmer world.

In CLIME, we paid particular attention to the uncertainties associated with the climate change projections. The number of climate simulations available has grown in recent years and most investigators now use a probabilistic, rather than a deterministic, approach to impact analysis (Samuelsson, Chapter 2, this volume). The schematic in Fig. 23.7 summarizes the approach adopted in CLIME to encapsulate this variability. The topics in the shaded boxes were not addressed by CLIME but must be considered together with all other risks if the objectives of the WFD are

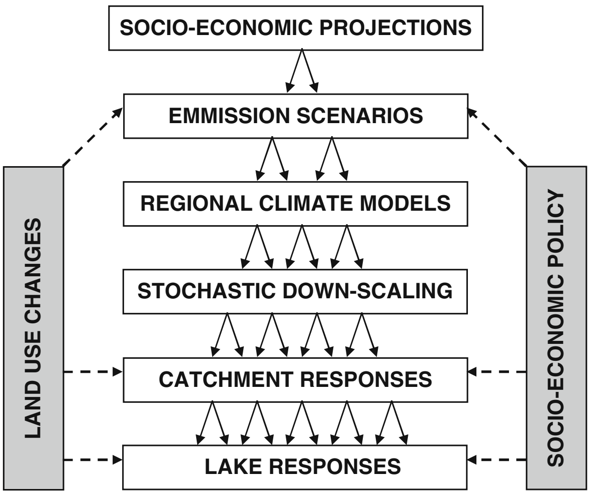

Fig. 23.7 The cascade of uncertainty associated with linking water quality models to the projected change in the climate (modified from Giorgi, 2005)

to be met in full. Although the number of climate models used to drive the CLIME models was limited, the projections are consistent with the range indicated in the most recent assessment produced by the IPCC. There is a danger that some managers will regard this as an opportunity to do nothing i.e. the 'wait and see' approach. It is now clear that the problems posed by climate change can only be resolved by a fully integrated approach to lake and catchment management. In the water industry, more attention is still being paid to the supply side rather than the demand side of the water quantity equation. Arnell and Delaney (2006) have described how the water industry in England and Wales is responding to the challenge. Here, the policy focus has moved away from impact assessments towards the development of adaptation measures. The general approach has recently been described in an EU Green Paper (SEC, 2007) and White Paper (SEC, 2009) on adapting to climate change. This recognises that water quality and quantity are intimately connected and that new pricing arrangements will have to be introduced to guarantee the efficient allocation of water. Our experiences in CLIME suggest that those charged with the implementation of the WFD are less aware of these institutional challenges. The Water Framework Directive provides an ideal means of addressing this issue in a holistic way but more needs to be done to integrate the disciplines, the data sets, and the expertise within the European research community.

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Chapter 24
Climate Change Impacts from a Water Supply Perspective

Lorraine L. Janus

24.1 Introduction

The primary aim of CLIME was to provide strategic scientific support for the Water Framework Directive (Chapter 23, this volume) but in addition to regional regulators, the project Advisory Board included delegates from a number of water suppliers. Water suppliers represent a different group of 'end-users' from the regional regulators and although they have the common goal of preserving water resources, they have different perspectives. Studies which form the basis of regulatory policy supported by the European Commission have focused on the broad impacts of climate change on water bodies throughout Europe. The report issued by the European Environment Agency on the impact of climate change on European water resources (EEA, 2007) is an example. Similarly, federal and state environmental conservation and public health agencies in the US have issued reports to describe potential national (Brekke et al., 2009) and regional (2001, Rosenzweig and Solecki, Ed.) climate change impacts. While these studies are of general interest in understanding how climate may affect regional issues, they do not address the more specific concerns of water supplies. From the water supply perspective, the challenge is to meet specific water quantity and quality goals (many of which are set by regulators) under the constraints set by unique watersheds and conveyance systems. Therefore, while regulators must define the goals for conservation and health, water suppliers must find practical ways to meet these goals.

This chapter looks at potential climate change impacts from a water supply perspective and describes how water supplies are likely to be affected by climate change. Water suppliers must maintain a reliable supply and meet regulatory standards despite large natural fluctuations in the environment. They typically devote much effort to developing structural and operational ways of buffering the naturally occurring extremes in water quantity and quality. Since extremes are predicted

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to increase in magnitude and frequency in the future, water suppliers will have to design adaptations to cope with these new conditions. An account of the way in which water companies in England and Wales are adapting to climate change was produced by Arnell and Delaney (2006). A comparable adaptation plan was issued by the New York City Department of Environmental Protection (NYCDEP) Climate Change Task Force (NYCDEP, 2008). These represent important first efforts in long-term planning needed by water supplies, but they are qualitative in nature, due to the large uncertainties in climate and hydrological predictions. The CLIME project is of particular importance from this perspective since it provides some innovative ways of addressing the uncertainties associated with even the most sophisticated climate change projections. The CLIME project uses these projections as drivers for water quality models to estimate how some important water quality parameters may change. These more specific predictions are what water suppliers need in order to evaluate their system's vulnerabilities and to plan appropriate adaptation measures for the future.

In CLIME, the problem of managing lakes and catchments in an uncertain world was addressed by combining the expertise of a group with much experience in integrated management with the analytical skills of the team assembled to quantify the limnological effects of future changes in the climate. The leaders of the catchment management team were based in New York where GWLF has been used extensively to identify and manage turbidity and nutrient sources for the New York City (NYC) water supply. Climatologists from the Swedish Meteorological and Hydrological Institute and the UK Climate Research Unit and systems analysts from the Helsinki University of Technology then played a key role in integrating the climate, lake, and water quality assessments. The decision support system produced by CLIME was based on a series of Bayesian networks informed by the climate change scenarios developed in Europe and adaptations of the catchment models used in the US. The problem of making investment decisions under uncertainty is not new, but climate change has introduced a new realm of application for DSS tools as demonstrated in the CLIME project.

In this chapter, a case-study approach is used to demonstrate the role that such a holistic approach can play in the development of water management strategies. The NYC water supply catchment was chosen for CLIME for both the extensive catchment modeling experience and the large, diverse reservoir systems thought to provide relevance for a broad audience of scientists and managers. In contrast to the situation in many metropolitan areas, the water distributed to consumers in NYC is of such good quality that it remains unfiltered with no treatment other than routine disinfection and fluoridation. The City's goal is to maintain this 'unfiltered' status through an anti-degradation policy that employs intensive watershed protection. The problems posed by the changing climate will only increase the difficulty of meeting this challenge. The chapter begins with a brief description of the current water supply system for NYC followed by a review of the climate change trends and projections for the NYC watershed. It goes on to discuss some of the water quantity and quality issues faced by managers, introduces some programs established to deal with these issues and explains how NYCDEP plans to cope with the projected impacts of climate change. Advance planning for climate change is very important for a large, complex system such as NYC's since policy-related research, and infrastructure design and construction can take decades to complete. Therefore decisions to invest in such projects must be made well in advance of their real need, despite uncertainty. The concluding section explains how NYCDEP's Climate Change Task Force (CCTF) is approaching this challenge and how this strategy has benefited from the collaborative studies undertaken by CLIME.

24.2 The NYC Water Supply System

The current NYC water supply consists primarily of three surface water systems known as the Croton, the Catskill and the Delaware Systems (Fig. 24.1). Parts of these systems were built over 150 years ago and the supply now consists of 19 reservoirs and 3 controlled lakes. As with all surface waters, the climate, hydrology, geologic setting, and land use are primary determinants of water quality in the reservoirs. Maintaining this system requires daily monitoring and active management by several hundred staff members to ensure delivery and compliance with all drinking water and environmental standards. These standards are defined in the Safe Drinking Water Act (SDWA) and the Clean Water Act (CWA), which govern water quality at the tap and throughout the watershed, respectively. The brief description given here outlines the evolution of the systems and highlights some key features. Further details can be found at the City's website (www.nyc.gov/dep).

The Croton System was the first of the City's three water systems to be built and came into operation in 1842. In subsequent years it was expanded and by 1911 included 12 reservoirs and 3 small lakes. It provides about 10% of the City's annual consumption. The watershed (catchment) of the Croton System is 970 km^2 and provides an average yield of 909 thousand cubic meters per day. The 12 reservoirs are typically mesotrophic to eutrophic and produce water of lower quality than the two later systems. The Croton catchments have higher population densities, include more wetlands and have, smaller and shallower reservoirs situated on more base-rich rocks, than the other systems. Water quality problems in the Croton System are typically those related to eutrophication, and include high algal concentrations, deep water anoxia and high levels of iron and manganese in the late summer. Organic matter generated by wetlands and algal production form disinfection by-products (DBPs) when the water is chlorinated and the Croton System is on the borderline of compliance with the current DBP rules. Although the Croton System is operated to meet all primary (health-based) standards, periodic violations of the secondary (aesthetic) standards for color, taste, and odor occur (see Section 24.4.2.2).

The Catskill System was the second to be constructed and was completed in 1926. It now provides about 40% of the City's annual consumption. This system consists of two reservoirs (Ashokan and Schoharie) with a watershed area of 1.48 thousand square kilometers that provide an average yield of 1.78 million cubic meters per day. These two catchments are located in the Catskill Mountains approximately 160 km away from the City. Land use in the Catskill and Delaware

Fig. 24.1 The Catskill, Delaware, and Croton System catchments of the New York City Water Supply

watersheds is 85% forest and undeveloped land. The geological history of the Catskill System is the main factor responsible for a recurrent water quality problem: extreme turbidity events. During the last ice age, the valleys that now form the stream corridors in these catchments were at the bottom of glacial lakes which accumulated layers of fine silt and clay. These clay deposits now form streambeds that are highly susceptible to erosion from intense rainfall and high flows. High stream flow in these large, mountainous catchments cuts through the clay deposits of the streambeds and results in very high levels of turbidity (i.e. 200–300 nephelometric turbidity units) in the reservoirs. Such turbidity events can persist for months at a time and represent the most serious water quality problem ecountered since the System was built in the 1920s.

The Delaware System was the third, and largest, system to be constructed and was completed in 1965. It now provides about 50% of the City's annual consumption. This system consists of four reservoirs (Neversink, Cannonsville, Pepacton, and Rondout) with a watershed area of 2.67 thousand square kilometers that provide an average yield of 2.2 million cubic meters per day. These catchments are located in the western and southern Catskill Mountains. The arrangement of the reservoirs in the Delaware System is such that three headwater reservoirs – Neversink, Cannonsville, and Pepacton – are connected via aqueducts that conduct flow into Rondout Reservoir. This allows operators to select the best water quality to convey to the City, with the potential to exclude water from an upstream reservoir if temporary impairment (such as high levels of algae, bacteria, or turbidity) occurs. The Delaware reservoirs are located on the western side of the Catskill Mountains and, as a result of different geology, are less susceptible to the extreme turbidity events than Catskill reservoirs. The Delaware System includes what is currently the most oligotrophic reservoir (Neversink) and the most eutrophic reservoir (Cannonsville) in the entire NYC water supply system. The primary anthropogenic impacts have been a few large and many small sewage treatment plants and more than 300 dairy farms, in catchments which are otherwise largely forested. These land uses have brought about the cultural eutrophication of Cannonsville Reservoir, which is expressed as chronic summer blooms of blue-green algae. Recent data suggests that the installation of tertiary treatment at all wastewater treatment plants in this catchment has been successful in reducing both the intensity and duration of these blooms. Today, the main focus of the watershed management program is the control of diffuse phosphorus inputs from dairy farms and stormwater. The Delaware System provides the largest share of the City's water supply and has received a proportionate investment in watershed programs.

The Catskill and Delaware Systems both flow eastward through two major aqueducts that reach a depth of more than 400 m below the Hudson River and resurface to the east. The difference in altitude between reservoirs and distribution is such that water will flow to the fourth floor of any building in the City without pumping, so the system is exceptionally energy-efficient. These aqueducts deliver water to the Kensico Reservoir, where the two systems converge; they are frequently referred to as the Catskill/Delaware System. The physical linkages between reservoirs are of great practical benefit since it allows the reservoirs to be used as a series of settling basins. As water moves through the system, suspended matter settles and the quality of the water improves as it moves downstream toward distribution. The network arrangement allows operators to exclude reservoirs with impaired water quality while selecting the best water for distribution.

24.3 The Climatic Changes Projected for the NYC Watershed Region

The Metro East Coast Report (Rosenzweig and Solecki, 2001) was one of the first publications to tackle the question of how climate may change in the northeastern United States. This report predicts significant changes for NYC and was instrumental in highlighting the need for an integrated planning process to guide adaptation to new conditions. The report reviews trends in climate over the past century and applies IPCC emissions scenarios and climate models to predict future conditions for the NYC region. Observations of changes over the past century (from 1901 to 2006) show annual increases in mean air temperature of around 1.5°C in Central

Park and 0.8°C in the watershed area. Annual precipitation increased by approximately 14.2 cm in Central Park and 7.4 cm in the watershed. Sea level rose approximately 30.5 cm over the past century; however, half of that can be attributed to subsidence. The City's Department of Environmental Protection has also obtained projections of future conditions, based on five models and three emissions scenarios (NYCDEP, 2008), as a preliminary basis for planning. Results suggest that the most probable changes for 2080 will be increases in air temperature of 3.9–4.4°C, a precipitation increase of 7.5–10%, and a sea level rise of 40.6–45.7 cm. Some additional predictions forecast increased intensity and frequency of both droughts and storms. Storms with a predicted recurrence interval of 100 years under current climate conditions are expected to occur every 10 years by the 2080s. An increase in the severity of storms combined with sea level rise will greatly increase the risk of significant damage in coastal cities like NYC.

In a study focusing on the City's most important water supply catchments, the US Geological Survey (USGS) compiled information on changes that have taken place in the Catskill Mountains over the past half-century (1952–2005). Based on 8–12 sites, annual mean air temperature has increased 0.6°C (with maximum daily air temperatures increasing the most over the winter period), regional mean precipitation by 13.6 cm (with a 7.6 cm increase in runoff), regional mean potential evapotranspiration by 1.9 cm while the peak snowmelt has shifted from early April to late March (Burns et al., 2007). In terms of water supply, this study indicates that more water is now supplied to the reservoirs earlier in the year, as a consequence of earlier snowmelt. At the same time, the observed increases in summer and fall precipitation and runoff suggest that reservoirs are likely to be full earlier in the spring, with potentially significant consequences for the annual water balance (see Section 24.4.1.2).

24.4 The Impact of Climate Change on Water Supply Management

The issues related to quantity and quality of water affected by climate change described below are examples from the NYC water supply catchments. The total area is very large (nearly 2,000 square miles) with significant geographic diversity, and contains 19 reservoirs that cover the full range of trophic conditions. The water supply management issues are diverse and are applicable to many other northern temperate locations, therefore the various aspects of watershed management explored below are relevant for many other surface water supplies in the northern temperate zone. The purpose of the discussion below is to provide insight into areas ripe for future research and planning related to climate change.

24.4.1 Water Quantity Management

It is clear that the City could not exist without the delivery of water in quantities sufficient to support the health and welfare of its residents. One of the challenges for the water supply is to provide a constant flow to meet consumer demand. In order to do this, the large natural fluctuations in hydrology between the extremes of floods and droughts must be buffered by having adequate storage capacity. If the quantity of water in storage is at maximum capacity, systems may become overwhelmed by floods, property damaged, and the quality of drinking water impaired. If the quantity of water in storage is at a minimum, basic functions such as sanitary sewers and firefighting capability could be affected. Maintaining the optimum amount of water in the system is a continuous process of evaluating the present and anticipating the future. This is further complicated by the fact that much of the City's water supply infrastructure is more than a century old and alterations or repairs to these structures are difficult. Beyond the difficulties of current-day maintenance, climate change will put new pressures on the City's aging infrastructure as they are challenged to cope with floods and droughts of greater magnitude. Recent extreme events have prompted NYCDEP to shift its operation strategy. The following sections describe approaches the NYCDEP and others have used to address the potential impacts of both current extremes and future changes in the climate on the quantity and quality of water delivered to consumers.

24.4.1.1 Floods

One of the first questions to be addressed in climate change planning is: how well will structures designed for current conditions withstand the floods of the future? The structures built by the Bureau of Water Supply include dams, bridges, and wastewater treatment works. All were designed to meet particular needs and to avoid risks associated with failure, thus they have very different lifespans. The federal Dam Inspection Act of 1972 mandated a Dam Rehabilitation Program to ensure that all dams throughout the United States meet specific engineering safety standards. Rehabilitation of the 19 NYC-owned dams began in the mid 1980s and is expected to conclude over the next decade at a total cost of more than $900 million. Dams are required to withstand the spillway design flood, which is based on the 'probable maximum flood' (PMF), defined as a storm with a probability in the order of 1 in 10,000 per year. The spillway design flood is generally 3–5 times greater than the most extreme flood of recorded. Given this very low probability of occurrence, global climate change is not expected to challenge the design flood and the required spillway capacity. The design criteria for smaller structures, however, are less robust and may well be affected by climate change. Bridges are designed to withstand storm events with a probability of 1 in 50 per year, and the individual characteristics of sewer systems make them susceptible to problems at even lower storm thresholds. Therefore, while the NYCDEP does not expect any problems with the dams, it may need to repair smaller public works at more frequent intervals. Climate change will also have to be taken into account when these smaller structures are replaced in order to minimize the risk of failure. One of the current projects recommended by the NYCDEP's Climate Change Task Force (CCTF) is the revision of the Infiltration, Duration, Frequency (IDF) curves used in the design of such structures (NYCDEP, 2008).

Operational considerations also have an influence on flood mitigation immediately downstream of the dams, and policies may need to be adjusted for future conditions. For example, the Delaware River Basin Commission formed in 1961 to mediate and integrate the management of water concerns and activities of New York City, New York State, Delaware, New Jersey, and Pennsylvania, recently adopted new, forward-looking policies to reduce the risk of flood damage. These policies were adopted in response to floods in 2004, 2005, and 2006 that ranked among the top eight flood peaks on record (DRBC, 2007; Suro and Firda, 2007; Lumia et al., 2006) and resulted in a 'state of emergency' and many millions of dollars of damage. The resolutions adopted by the Commission included: (i) controlled releases from the City's Delaware System reservoirs when combined storage capacity is over 75%, (ii) establishment of a task force to recommend flood mitigation measures, (iii) development of a model to evaluate future operational alternatives and (iv) installation of two snowpack gauges as part of the National Oceanic and Atmospheric Administration (NOAA) Automated Flood Warning System. Model development and climate change were specifically identified as considerations in the context of planning future flood mitigation. The NYCDEP is currently developing an Operational Support Tool (OST) that employs the 'Options AnalysiS in Irrigation Systems' (OASIS; see www.iwmi.cgiar.org) model to monitor the water balance of this large, complicated system. (It will guide current operations and can also be used to evaluate a range of different watershed management options. The OST will be important to the NYCDEP in improving real-time system operations to benefit both the supply needs of the City and addressing the habitat concerns of downstream stakeholders. As operating policies become more complex to accommodate competing needs, more reliable predictive models will be required to inform and guide future operations and policies.

24.4.1.2 Droughts

Drought occurs when supply is insufficient to meet demand over an extended time period. Droughts can be natural events caused by long periods of minimal precipitation, or man-made events known as 'operational' droughts e.g. when major segments of the system are taken out of operation for repair. Droughts are classified in stages according to their severity where each stage is defined by the probability of refill by June 1st. The stages of severity are Drought-Watch, Drought-Warning, and Drought-Emergency. Each stage triggers specific and increasingly stringent measures of conservation as specified in NYCDEP's 'Drought Management Plan and Rules'. High-demand activities (such as lawn watering and car washing) may be prohibited, and more time is devoted to educating the public on conservation practices.

There are two basic approaches to managing drought - supply enhancement (i.e., addition of new sources or increased storage) and demand reduction (conservation). Since the 1990s, the NYCDEP has been very successful in decreasing demand from an average of 5.7 million cubic meters per day to about 4.9 million cubic meters per day through programs providing leak detection and repair, fire hydrant locks, low-flow toilet rebates, and conservation education. Although this reduction in demand is a significant achievement, future conditions of increased population and temperature will tend to negate these historical gains.

In 2006, the NYCDEP initiated a Dependability Study to address the issue of shortages that can be caused by infrastructure maintenance, repair, and new future demands. The objective of this study is to ensure that there is a sufficient supply of water to sustain the demands of approximately ten million consumers under both natural and operational droughts. The Dependability Study will evaluate a wide variety of options to both enhance the supply and reduce demand wherever possible. The options under study include: additional demand-reduction incentives, maximizing the use of existing reservoirs, increasing storage capacity of existing reservoirs, repairing aqueduct leakage, building interconnections with other utilities, aquifer storage and recovery, desalination of harbor and ocean water, improvements to the Hudson River intake, and the construction of new water tunnels. Some of these options can substantially enhance capacity. Repair of a known leak in the Delaware Aqueduct will reclaim approximately 130 million liters per day that is currently lost from the system. Hudson River withdrawal at the Chelsea Pump Station is an option already in place that can supply up to 380 million liters per day. Although this intake was last used during a severe drought in 1989, its future use would present some practical problems. The Hudson River withdrawal is subject to very stringent monitoring and treatment in order to meet standards for an unfiltered supply, and to guard against introduction of invasive species, such as zebra mussels. In addition, the river contains sediments contaminated with polychlorinated biphenyls (PCBs) which are unacceptable to consumers even in trace amounts. In the future, if droughts became more severe, the salt front from the Atlantic Ocean would extend so far upstream as to render this intake unusable whenever there was not enough flow in the river to push ocean water downstream. The on-going sea level rise of about 25 cm per century (Rosenzweig and Solecki, 2001) would exacerbate the problem. If Hudson River withdrawal becomes critical, one potential consideration would be a desalination and treatment facility at the Pump Station.

The Dependability Study explores many options for supply management, enhancement, and conservation and their combined effects must be considered in selecting the most effective combination. If droughts become more frequent, the improvement alternatives identified by this study will become more valuable. However, these alternatives should still be reviewed at regular intervals in light of the most recent climate change scenarios since these new conditions could well influence the relative costs and benefits of the alternatives and may reveal the need for new technical solutions and infrastructure.

Management plans will also have to take full account of the natural factors that have a significant impact on the overall water balance of the watershed. Climate change will affect both the snowpack and the vegetation, which will have significant implications for the timing and seasonality of streamflow. The USGS study of the Catskill region (Burns et al., 2007) suggests that peak snowmelt has shifted to two weeks earlier in the spring than 50 years ago. If this trend continues, the change in the amount and timing of snowmelt entering the reservoirs will have a major effect on drought management, the onset and duration of stratification, and managed release of cold water for downstream fisheries. Similarly, any major climate-related change in the vegetation would also significantly affect the water balance. At present, the proportion of water lost by evapotranspiration accounts for about 40% of the total annual precipitation (Murdoch, 1991; Randall, 1996). As vegetation changes, so will the overall water balance, and in particular, streamflow. Over the long term, models used to track mass water movements will need to account for differences in snowpack, land use, and forest assemblages as they evolve in response to climate.

24.4.2 Water Quality Management

The management programs implemented by the City to comply with current drinking water regulations are designed to address four main issues: (i) bacteria and pathogens, (ii) organic precursors to disinfection by-products (DBPs) and other dissolved substances, (iii) phosphorus loads, and (iv) episodes of very high turbidity. In the following sections, these issues are discussed, in the order of priority suggested in an extensive review of the City's watershed management program (National Research Council, 2000). Climate change will undoubtedly have an impact on all of these water quality issues, and on the efficacy of the programs designed to control them.

24.4.2.1 Bacteria and Pathogens

Management of bacteria and pathogens in the water supply is of particular importance for the NYCDEP to maintain its filtration avoidance status. Kensico Reservoir is the 'source water' for both the Catskill and Delaware Systems and is therefore subject to the fecal coliform standard and *Giardia* monitoring specified in the Surface Water Treatment Rule (SWTR) of 1989. Since studies at this site have shown that waterfowl and stormwater are the primary sources of fecal coliform bacteria in the reservoir (NYCDEP, 1993), programs have been introduced to reduce the bacterial loads from these sources (i.e. in European terminology, programs of measures have been introduced). In 1998, the SWTR was amended to include regulation of the pathogenic protozoan *Cryptosporidium*, so programs for monitoring and management of this pathogen have also been developed.

The control of fecal coliform bacteria in the City's most important source water is achieved through a Waterfowl Management Program (WMP), which has proven to be exceptionally effective. This program was developed as a result of a study completed in 1993, and was based on a bacteriological mass balance, coliphage identification, and ribotyping (NYCDEP, 1993). The study identified waterfowl as the primary source of fecal coliform bacteria at the intakes. The result was development of a waterfowl deterrent program, which demonstrated that when waterfowl were prevented from roosting on the reservoir at night, coliform bacteria levels were reduced to well below the regulatory limit. The results in (Fig. 24.2) demonstrate

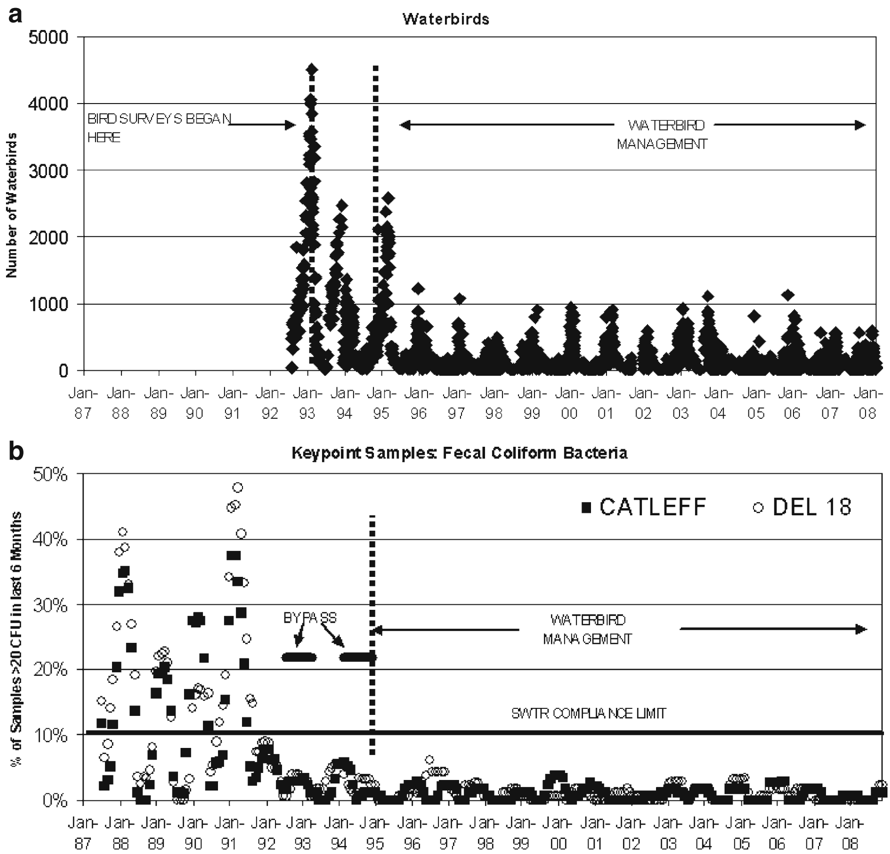

Fig. 24.2 Effectiveness of NYCDEP's Waterfowl Management Program for reducing fecal coliform bacteria at Kensico Reservoir: (**a**) The change in waterbird population between 1992 and 2008. (**b**) Fecal coliform bacteria at reservoir intakes (DEL18 and CATLEF) between 1987 and 2008. The *horizontal line* shows Surface Water Treatment Rule (SWTR) compliance limit

the effect of the WMP. The upper panel of the figure shows the number of birds on the reservoir since surveys began in 1991 and demonstrates that these numbers declined sharply after 1993 when waterfowl management (i.e. chasing birds from the reservoir and addling eggs) began. The lower panel shows the direct response of coliform bacteria measured at the intakes over the same time period. The direct correspondence between the presence of waterfowl and the coliform levels has been demonstrated when the program ceased for two weeks and coliform levels began to increase. This program has worked exceptionally well since its implementation and has eliminated the need to bypass the reservoir, which is an alternative way to exclude reservoir water when impaired. In recent years, Kensico Reservoir has consistently remained well below the regulatory limits for coliforms. The WMP has been in operation since 1993 and will undoubtedly continue indefinitely into the future because of its critical role in maintaining filtration avoidance status.

As the climate warms, the migratory patterns of the birds may well change and the current seasonal populations could become full-time residents. Greater availability of open water for roosting during the winter, coupled with a longer growing season for foraging could prove to be important. As a result, the problem may change from one that is primarily associated with fall and winter migration to one that continues throughout the year. The current program, which operates at a cost of more than $1 million per year, would then have to increase in duration and greatly increase the costs of control.

Stormwater is another way in which bacteria and pathogen cysts enter the water supply. At Kensico Reservoir, stormwater inputs are buffered by control structures. The City installed 45 stormwater control structures, including a variety of detention and infiltration basins to provide pre-treatment of the water before it enters the reservoir. Monitoring has demonstrated that these structures are effective in reducing concentrations of fecal coliform bacteria, suspended solids, and total phosphorus by about 50% for storm flows that are within the design capacity of these controls. The structures are designed to retain the first flush of approximately 3.8 cm precipitation events in 48 h (i.e., storms with a return frequency of about 2 years). The projected increases in the rainfall imply that, many of these basins may need to be enlarged or modified to remain fully effective. In addition, as rainfall and runoff change, the statistics and coefficients used in the design of these structures will have to be revised to be effective under new storm flow conditions.

The City's pathogen program has clearly demonstrated that concentrations of *Cryptosporidium*, *Giardia*, and enteric viruses are currently at very low levels and similar to those found in pristine watersheds. Unfortunately, the specified monitoring procedures do not distinguish between different species of pathogen genera, and as a consequence, all species, including those which are not infective for humans, are included. Microbiological fingerprinting techniques (genotyping and ribotyping) have, however, shown that the *Cryptosporidium* and *Giardia* species found in the water supply primarily come from the local wildlife, mostly deer, raccoons, skunks, and mice, and are not considered primary human pathogens. In 2007, the City submitted extensive historical data (as required by the Long-Term 2 Enhanced Surface Water Treatment Rule) to the EPA and now expects to be classified in the lowest rank for pathogen risk. Current watershed protection programs are designed to further reduce this risk by minimizing all potential sources of pathogens and their transport pathways.

Cryptosporidium is resistant to chlorine disinfection, so managing this threat to the water supply involves limiting it at its source. In the Catskill/Delaware watersheds, home to nearly 300 dairy farms, one such potential source is cows. A main goal of the Watershed Agricultural Program, therefore, is to promote practices that limit the spread of *Cryptosporidium* from these animals to the natural flow-path of water. One practice that seeks to achieve this is the separation of calves into individual greenhouses to eliminate transmission of the disease through close contact between the young animals. Other management practices include: proper composting of manure to reduce pathogens, correction of drainage problems, restrictions on the spreading of manure when the ground is frozen or near watercourses, maintenance of the vegetative buffers along the watercourses, and the control of cattle feed quality to reduce the amount of phosphorus in manure (Cerosaletti et al., 2004). Much of the success in implementing these practices is due to the fact that this is a voluntary program. It is administered by the Watershed Agricultural Council, funded by the NYCDEP and supported by 90% of the farms in the watershed. With climate change, the primary questions will revolve around changes in hydrology and whether or not drainage systems are sufficient to keep these sources of pathogens and nutrients away from natural flow pathways. Structural capacities, buffer distances, and drainage patterns will all have to be monitored and periodically evaluated to ensure that new hydrological regimes do not outstrip the capacity of these measures to remain effective.

Giardia is another pathogenic protozoan that is a potential cause of waterborne disease. This organism is less of a threat than *Cryptosporidium* because it is susceptible to disinfection by chlorine. However, in the context of climate change, it is of interest because it displays a distinct seasonality, where the numbers of cysts in the water supply (Fig. 24.3) are inversely correlated with temperature. Several years of observations of different reservoirs demonstrate that concentrations of *Giardia* cysts are higher during the winter than during the summer. At this time, it is not known why this occurs, but a number of possible explanations have been suggested. These include lower settling rates during the winter due to the increased density of cold water, greatly reduced numbers of filter-feeders that would tend to remove cysts, decreased natural UV disinfection in winter, and seasonal changes in the behavior of the host mammals. The observed seasonality of *Giardia* certainly suggests that

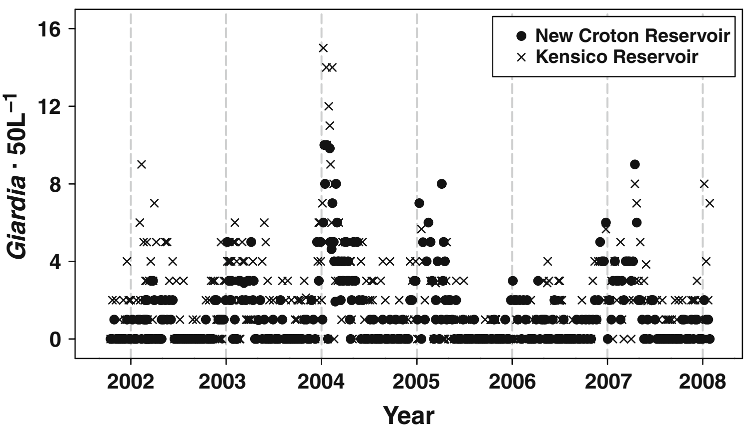

Fig. 24.3 Seasonality pattern of Giardia cysts at Kensico and New Croton Reservoirs from 2002 to 2008, with *vertical lines* indicating January of each year

there are many ways in which the ecology of the organism and its hosts could be significantly affected by temperature. Further study of the factors responsible for the observed correlation with temperature will be needed before we can predict this pathogen's response to the projected changes in the climate.

Another factor that appears to influence pathogen cysts is the concentration of suspended particles, in stormwater. High concentrations of suspended particles may enhance the sedimentation rates of pathogen cysts by coalescing with them and accelerating the rate at which they sink out of the water column. Observations on pathogens during a major storm and the resulting turbidity event in 1996 are consistent with this idea. Almost no pathogen cysts were observed during this storm of rain on snow and no peaks occurred despite the initial mass mobilization of small particulates and other substances. The increased frequency of extreme weather events projected in a warm world could well affect the sedimentation dynamics of pathogen cysts. As our understanding of the interaction between particles and pathogens improves so too will the design objectives of stormwater control structures, pathogen transport models, and the evaluation of risks of pathogen contamination associated with storms.

24.4.2.2 Organic Precursors of Disinfection Byproducts and Dissolved Substances

Organic dissolved substances in water supply reservoirs are important because of their potential impacts on human health and on the aesthetic quality of drinking water. Dissolved organic carbon (DOC) is of particular concern because it can form carcinogenic compounds when water is disinfected by chlorination. The compounds that form are known as disinfection byproducts (DBPs) and are considered a health risk when they reach certain concentrations. Other dissolved organic compounds, such as those generated by phytoplankton can cause taste and odor problems and in some cases, can even be toxic. Among the inorganic dissolved substances that are problematic are iron and manganese, which impart a dark color to water that diminishes its appearance and acceptance by consumers. The origins of these dissolved organic and inorganic substances are primarily wetlands, phytoplankton, and sediments. An understanding of their origins and conditions related to climate that promote their production is key to their current and future management.

Under current conditions, the Catskill/Delaware reservoirs contain very low concentrations of carbon. Total organic carbon (TOC) in these reservoirs averages 1.6–2.2 mg L^{-1}. Organic carbon from wetlands is minimal since they represent only 1% of the surface area of the Catskill catchments. Because of the small percentage of wetlands, climate change will affect dissolved substances in these reservoirs largely through its impact on primary productivity, which is currently low because of the oligo- and mesotrophic nature of the reservoirs. This could increase as changes in temperature and hydrology occur, and.therefore, the NYCDEP has already taken an initial step to evaluate the potential for increased primary production by applying the GWLF and the PROTECH models to Cannonsville Reservoir. These models were the ones used in CLIME and have been described in Chapters 2 and 15 of this volume.

In contrast, Croton System water quality suffers from both external and internal sources of dissolved substances. Croton reservoirs range on average from 3.2 to 5.2 mg L^{-1}TOC (with most of this present as DOC) which is nearly twice the concentration of TOC as the Catskill/Delaware reservoirs. The Croton catchments are approximately 6% wetlands and studies have shown that most DBP precursors in the Croton system originate in wetlands (NYCDEP, 2003). In addition to the external sources of dissolved carbon, New Croton Reservoir is classified as eutrophic. Therefore, phytoplankton and macrophyte production add to the carbon compounds from wetland sources. As a result, the current concentration of haloacetic acids (a class of DBPs collectively referred to as HAA5) in the Croton System is very close to the permitted maximum and must be managed to avoid exceeding the drinking water standards.

Dissolved inorganic substances are also an issue. The hypolimnion typically becomes anoxic in August, which results in the release of iron and manganese from the sediments. Dissolved iron and manganese impart a dark color to the affected water that causes it to exceed the secondary (aesthetic) standard. In recent years, the Croton intakes have been managed by simply shutting them down during the late summer months in order to ensure compliance with objective water quality standards and to avoid sending highly colored water into distribution.

In future decades, the projected changes in temperature and precipitation will accentuate the problems posed by dissolved substances in the Croton System by altering both the. trophic status of the reservoirs and wetland density. Management of these water quality problems may require additional treatment for the compounds' removal as well as careful adjustment of the levels of chlorination (NYCDEP, 2004). Levels will have to be optimized to retain residual chlorine in the distribution system while at the same time minimizing the rate of byproduct formation. A filtration plant for the Croton System is scheduled for completion in 2012. While it will reduce the risk of microbial contamination by removing protozoans, bacteria, plankton, and other particulates, it will not resolve all the problems arising from increases in dissolved constituents. The problems include precipitation of dissolved iron and manganese (from anoxic sediments) on filters and the potential need to pretreat water for removal of organic DBP precursors. Once the plant is in operation, any increases in these substances brought about by climate induced eutrophication may diminish its capacity and add to the costs associated with the operation of the plant.

24.4.2.3 Phosphorus

Eutrophication and high levels of algal production can have many impacts on water quality, including taste and odor problems, organic precursors of DBPs, oxygen depeletion, dark color, and fish kills. The trophic condition of waterbodies is to a large extent determined by phosphorus (Vollenweider and Kerekes, 1982).

Accordingly, target concentrations for phosphorus have been set at 20 mg.m^{-3} for upstream reservoirs and 15 mg.m^{-3} for the reservoirs close to distribution. In order to meet these targets, Total Maximum Daily Loads (TMDLs) for phosphorus have been defined for each reservoir. Phosphorus loads are typically derived from both point sources (e.g., wastewater treatment plants) and nonpoint sources (e.g., agricultural fertilizers, septic systems, etc.). A major step in meeting the phosphorus loading goals was made in recent years under the City's Watershed Protection Program. Here, approximately 100 wastewater treatment plants in the watershed were upgraded to tertiary treatment with microfiltration or its equivalent. As a result, the phosphorus load targets have been met for all three watershed systems. Nutrient enrichment from nonpoint sources, however, remains a problem in some basins, and although these problems are currently being addressed (see, e.g., Section 24.4.2.1), the projected increases in the frequency of extreme weather events will almost certainly lead to an increase in the mobilization of nutrients and the rate at which these are transferred from the landscape to the reservoirs. The trade-off between increased loading and increased flushing rate will ultimately determine the new trophic status of the reservoirs (Vollenweider, 1976). Simulation models, such as those described in Chapters 9, 11 and 15 can be used to evaluate the different control measures required to mitigate the effects of the projected changes in the climate.

24.4.2.4 Turbidity

Turbidity is considered a significant threat to water supplies because it can interfere with disinfection, introduce undesirable contaminants, or prevent water from meeting the prescribed aesthetic standards. Unacceptable levels of turbidity in the Catskill System (due to storms) are typically controlled through operational adjustments, and as a last resort, alum treatment at the downstream end, where alum is added to the Catskill Aqueduct just prior to entry into Kensico Reservoir. As the treated water enters the reservoir, the flocculated clay and alum settles rapidly to the bottom where it remains until it is removed by dredging. At present, these alum treatments are sufficient to reduce turbidity to a level that meets the regulatory standard. In the case of NYC's water supply, violation of the turbidity limits at the raw water intakes could lead to loss of filtration avoidance status and a regulatory order to construct a filtration facility. Such a facility would have to be capable of handling peak flows of up to 7.5 million cubic meters per day and would cost on the order of six to eight billion dollars. Control of turbidity, therefore, is fundamental if the City is to avoid costs of this magnitude.

The primary cause of these turbidity events is unusually high or intense rainfall. The Catskill System is particularly susceptible to turbidity events due to its past geological history. The swift stream flows that result from heavy rains erode the layers of glacial clay that cover the region and turn the streams and reservoirs a muddy, reddish-brown. The work to date suggests that the erosion of the streambeds is the primary source of this sediment, rather than the surrounding landscape (Nagle et al., 2007). Any increase in the intensity of rainfall as a result of climate change, therefore, will likely heighten turbidity peaks in the streams and reservoirs because of the increase in streamflow that accompanies greater rainfall intensity.

Temperature is another factor that plays a key role in modifying the severity and duration of turbidity events, through its effects on both landscape and reservoir processes. Turbidity events vary in severity according to the season in which they occur. Storms during the early spring, late fall and winter (i.e., leaf-off seasons) have a greater impact than those during the summer. In the summer, soils are permeable and a significant proportion of the water percolating through the soil is transpired by the growing vegetation. Rates of runoff are consequently relatively low which in turn decrease streamflow and the attendant mobilization of suspended particulates. Once the particles carried by the streams reach the reservoirs, their rates and dispersion and settling are also influenced by the thermal characteristics of the water column. Settling rates tend to increase as the water temperature rises but the effect of this change could well be offset if mixing events are either more frequent or more intense.

The condition of the surrounding forest is another factor that can have a large impact on turbidity. Forests represent the dominant land cover type in the Catskill and Delaware catchments and the changes in temperature and moisture associated with climate change may increase the success of fungal diseases or insect populations that can destroy forests. Loss of forests can destabilize soils in the watershed, leading to erosion and turbidity. In a recent case in Colorado, an infestation of mountain pine and spruce beetles destroyed much of the forest in the headwater catchments of the water supply for the City of Boulder. This left the forest susceptible to wildfires and severe erosion, which has subsequently resulted in increased turbidity in the water supply (Miller and Yates, 2006). Forest decline, whether through disease or logging, can also have an impact on water chemistry via the release of nutrients from soil into ground and surface waters. The USGS has studied nutrient dynamics under various conditions of logging to explore the effects of de-forestation on the transfer of nutrients from the soil to groundwater (McHale et.al., 2007). Changes in the interactions between rainfall, temperature, vegetation, and soils caused by climate change will have ecological consequences that affect both water chemistry and turbidity. More terrestrial modeling work clearly needs to be done to quantify the many factors involved in the mobilization and transport of the particulates that result in turbidity. These should be linked to water quality models that can guide decisions on aqueduct flows and the types of treatment required to make sure that the downstream intakes meet the established turbidity standard (<5 NTU). The changing nature of these factors makes it important to continue the work of water quality monitoring and to recalibrate the existing turbidity models as new information becomes available. In order to perform well, models must evolve along with the environmental changes that they are intended to portray and must be extended to reflect new processes.

24.5 NYCDEP's Approach to Climate Change

Changes in the climate, on the scale now projected have the potential to create significant problems for the management of water and wastewater systems in the City. The City's existence depends on these systems to function continuously through extreme climate conditions. We have some knowledge of the disruption and damage that such malfunctioning can cause through past experiences of extreme events. However, these experiences only provide a qualitative indication of what can be expected in the future. Our participation in the CLIME project, coupled with work already underway, allowed us to refine existing models to reflect the uncertainties associated with current climate change scenarios. As the models are developed, the impacts on water systems can be better evaluated to determine what new infrastructure and policies are needed to sustain the current level of service under future climatic conditions. Despite the uncertainties of the current science, long-term planning is already underway with all current systems and procedures re-evaluated in the context of sustained climate change. Both the CLIME Project (see Section 24.6) and the NYCDEP's Climate Change Task Force (CCTF) have contributed significantly to the development of the methods used for long-term planning.

The primary planning work related to climate change for the City water supply and wastewater is done through the NYCDEP's CCTF. The stated mission of the Task Force is to 'ensure that NYCDEP's strategic and capital planning take into account the potential risks of climate change - sea level rise, higher temperature, increases in extreme events, changes in drought and flood frequency and intensity, and changing precipitation patterns on the City's water supply, drainage, and wastewater management systems'. To accomplish this mission, the CCTF established a government-university collaboration with Columbia University's Center for Climate Systems Research (CCSR) and the NASA Goddard Institute of Space Studies (GISS). This collaboration gave the CCTF access to future climate scenarios (Rosenzweig et al., 2006) to use as a framework for evaluation and long-term planning for adaptation and mitigation. The evaluation of water system vulnerabilities and planning recommendations were consolidated in the report entitled: Assessment and Action Plan – A Report Based Upon the Ongoing Work of the DEP Climate Change Task Force (NYCDEP 2008), which can be found at http//www.nyc.gov/dep.

To ensure that climate change is considered in its strategic and capital plans, the NYCDEP is in the process of developing an adaptation strategy that sets forth approaches for gradually replacing, improving, or adding infrastructure to meet new needs. Gradual adaptation is important for the City in order to allow enough lead-time for large construction projects and to reduce the associated financial risks. The capital plan to support the water systems over the next decade (2005–2015) is $16.2 billion US (NYCMWFA, 2006). Long-term planning on this scale is necessary to protect the City's investments and to ensure that the added risks due to climate change are considered in infrastructure design, site selection, and new policies. The climate scenarios used for these assessments will be periodically updated using the cyclical process described in the Assessment and Action Plan (NYCDEP,

2008) noted above. The proposed assessment and adaptation process consists of eleven clearly-defined steps as follows: (1) identify the climate change impact to NYCDEP's water supply, drainage and wastewater management systems, (2) quantify the impact using watershed, sewershed, and harbor water quality models, (3) evaluate the likelihood that the impact will occur, (4) estimate the timeframe within which the impact will become significant, (5) identify various types of adaptation strategies to overcome or minimize the impact, (6) quantify the cost and effective lifetime of each adaptation strategy, (7) evaluate the risks associated with proceeding or not proceeding with the adaptations by comparing the estimated impact, the likelihood of occurrence, and the economic, engineering, and environmental aspects of the strategy, (8) develop a financial model that will sustain the climate change investment, (9) develop indicators that would trigger the need to start implementation of the adaptations, (10) implement and monitor the success of adaptations, and (11) track advances in climate modeling and periodically re-evaluate the need for additional adaptations. Systematic scrutiny of projects in this way should ensure that investments are made wisely. Climate change planning, which, as noted above, requires decision-making on some very large and costly construction projects, can be informed through the application of Decision Analyses which can also be used to quantify the associated uncertainties. This Bayesian technique is a systematic way of incorporating uncertainties and attitudes about risk and is therefore well suited to climate change planning. The CLIME-DSS, described in Chapter 22, showed what can be achieved when data from long-term observations and simulation modeling can be combined and presented in a user-friendly way. Our experience of working with these system analysts proved most instructive and comparable systems are currently being planned for NYC. Beyond the conclusions reached for water quantity and quality changes, this technique will be applied to assist with decisions on policies, infrastructure, and optimal timing for implementation of these adaptive measures.

A number of other mitigation and litigation actions taken by the City illustrate its progressive approach to climate change. The City has taken a proactive role in the mitigation of global warming by reducing its greenhouse gas (GHG) emissions and pledging a 30% reduction by 2030. It has also begun to inventory GHG emissions using the software supplied by the International Council on Local Environmental Initiatives (ICLEI). The current inventory demonstrates that approximately 80% of the NYCDEP's emissions from all water and wastewater operations come from eight sludge de-watering facilities. These facilities will be targeted for GHG reductions while many of its existing gas-powered vehicles will be replaced by hybrid (gas-electric) equivalents. Litigation in progress includes legal challenges to both the US Environmental Protection Agency, for failure to regulate gas emissions, and the National Highway Traffic Safety Administration, for improper environmental review of the rules used to classify vehicles as 'light trucks'. In another case, the City joined the states of California, Connecticut, Iowa, New Jersey, New York, Rhode Island, Vermont, and Wisconsin to force the five largest US power plant emitters of carbon dioxide to gradually reduce their emissions even in the absence of federally-mandated standards (*State of Connecticut* et al. *vs. AEP* et al., Civ. No. 04-5669

(SDNY 2004)). The energy and cost savings associated with actions like these have greatly accelerated the trend toward GHG reductions. Moreover, New York is not alone in taking such actions – many cities and companies throughout the nation are involved in similar actions to minimize the projected effects of GHG emissions. Further information about the City's long term planning can be found on the internet under PlaNYC.

24.6 The Importance of the CLIME Collaboration

The European Commission has always recognized the value of scientific collaboration between its Member States and specialists from other countries. CLIME was one of the first Climate and Environment projects to integrate a study planned in Europe with work underway in the US. From the outset, the plan was to adapt some established methods to quantify the potential effects of climate change on the management of lakes in a range of different environments, and through collaboration, CLIME has succeeded in doing this.

Collaborations of the kind described here, are likely to become increasingly common as the range of issues covered by these European research programs continue to expand. In 2007, the European Union launched its Seventh Framework Program (FP7) for Research and Development which covers the period between 2007 and 2013. The proposed programs should be both easier and simpler to work with than earlier EU programs and new procedures have been introduced to encourage more collaboration with the US. In the case of CLIME, the NYCDEP was the only US partner in a consortium of 17. Researchers, universities, institutes and commercial firms established in the United States are all allowed to participate in FP7 and a copy of the Rules for Participation, as published in the EU Official Journal, can be found on: http://eur-lex.europa.eu/LexUriServ/site/en/oj/2006/l_391/l_39120061230en00010018.

Since CLIME was a three-year project, it was clear that the models used to simulate the responses of lakes to the projected changes in the climate had to be at an advanced stage of development. The GWLF model as developed by the City met these criteria and was also supported by several years of on-site calibration and testing. The approach adopted by CLIME to promote the widespread use of these models proved particularly effective. In this respect, the City's modeling staff played a key role in translating the model into the more user friendly VENSYM environment which greatly simplified the task of extending the scope of the models and applying them to an exceptionally wide range of climatic conditions.

Similarly, the research group from New York was able to adapt the procedures used to downscale the climate change scenarios to meet the needs of their own modelers. The methods used to assess the impact of climate change on natural systems are currently undergoing a shift away from deterministic techniques to stochastic procedures that encapsulate the uncertainties associated with these projections. The scenarios used within CLIME were confined to the European area but the same techniques are now being applied for the NYC region in an attempt to produce climate scenarios that are more relevant for the watershed area. Both organizations benefited from the experiences gained by running the climate, catchment and lake models in sequential mode, i.e. where the outputs from one model are used as the inputs to the next model in the chain. The geographic spread of watersheds included in CLIME also provided new insights into the climatic sensitivity of lakes and how their responses might change under warmer conditions.

From a practical point of view, the most innovative aspect of the project was the development of the Decision Support System (CLIME-DSS). A key issue now being addressed by watershed managers is how existing programs of management might have to change in order to remain effective under future climate conditions. The CLIME-DSS described in Chapter 22 provides the water managers in Europe with the information they need to explore the potential effect of climate change on the physical, chemical and biological characteristics of the lakes and their surrounding catchments. Steps are now being taken to develop this approach for climate change planning in the US. The uncertain nature of climate change and the large risks associated with it have renewed interest in the application of Bayesian networks as a means of integrating the results of different climate and water quality models to quantify the associated risks. The US Climate Change Science Program (CCSP) is currently developing DSS tools for planning, decision-making, and policy-making that can accommodate these uncertainties and explore the consequences of abrupt climate changes (http://www.climatescience.gov/Library/stratplan2008/). The application of DSS is an area of applied mathematics that will undoubtedly flourish as the pressure to make costly decisions about an uncertain future increases. The DSS application in CLIME represents pioneering work in the development of ways to integrate science from various disciplines to guide water resources policy and planning for the future.

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Index

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