Global Environmental Studies

Makoto Taniguchi
Tetsuya Hiyama *Editors*

Groundwater as a Key for Adaptation to Changing Climate and Society

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Groundwater as a Key for Adaptation to Changing Climate and Society

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Foreword

The impacts of climate change will be felt mainly through impacts on the water cycle, driven by changes in regional weather patterns, particularly rainfall and extreme events. While much of the scientific and public attention has been focusing on the implications of climate change on surface water resources, little regard is being paid to the lion's share of the unfrozen freshwater resources on our planet: groundwater.

Almost half of the world's population depends on groundwater as its primary source of drinking water. Supplying about one third of the world's irrigated land, groundwater is critical to global food security and in many arid and semi-arid areas groundwater is the only reliably available source of freshwater. Many ecosystems, hosting rich biodiversity and providing valuable services to humans, are fed by groundwater. In light of global changes including population growth, urbanization, land-use changes, economic development, and climate change, the pressure on this resource is steadily increasing: not only are a growing number of people dependent on groundwater for their daily needs, but also contamination is threatening the high quality of groundwater resources in many areas.

Climate change, in conjunction with human activity, is expected to have severe impacts on both the quality and quantity of groundwater resources globally. While recognizing the importance of groundwater in supporting human livelihoods and ecosystems, the Intergovernmental Panel on Climate Change (IPCC), in both its third and fourth Assessment Reports, states that "there has been very little research on the impact of climate change on groundwater. . .." The IPCC Technical Paper on Climate Change and Water (2008) remarks that "information about the water-related impacts of climate change is incomplete, especially with respect to water quality, aquatic ecosystems, and groundwater, including their socio-economic dimensions".

Addressing the limited knowledge and research related to the impacts of climate change and human activity on groundwater resources, UNESCO's International Hydrological Programme (IHP) initiated the GRAPHIC project (Groundwater Resources Assessment under the Pressures of Humanity and Climate Change) in 2004. The project provides a platform for communication and exchange among groundwater and climate experts around the world. It aims at improving our understanding of how groundwater interacts within the global water cycle, how it supports ecosystems and humans, and in turn, responds to the combined pressures of human activity and climate change.

The physical impacts of climate change and human activity on groundwater will vary greatly depending on geographical location, climate scenarios, and subsurface conditions. In order to better understand current processes and to allow for the prediction of future changes, case studies should be used to investigate how climate change impacts on groundwater resources under specific climatic and hydrogeological conditions. In addition, we also need to be evaluating the social and economic implications on society. Within the framework of GRAPHIC, numerous case studies have been conducted over past years in Africa, Asia and the Pacific, Europe, and North America, as well as in Latin America and the Caribbean, covering a broad range of geographical and climatic settings. In order to address the complexity of multisystem interactions, GRAPHIC incorporates a multidisciplinary scientific approach. Recommendations based on the scientific research and relevant for policy and decision makers are derived from each case study. Bringing these to the attention of the global community of water managers, scientists, and politicians is one of the main objectives of the project.

Despite being far from complete, the knowledge of climate change impacts on groundwater has increased significantly. Over the past two decades, there has been a steady rise in scientific publications on the subject. While only a small group of scientists conducted research on climate change impacts on groundwater resources up until the mid-1990s, the number of scientific publications including both "groundwater" and "climate change" as keywords rose to 50 publications per year in 1999. Since then, groundwater and climate change-related publications have been steadily increasing, reaching more than 250 publications by 2009, according to a query in the SciVerse SCOPUS database (http://www.info.sciverse.com/scopus/).

However, a better understanding of the role of groundwater in the context of climate change does not automatically lead to higher recognition of the subject among those responsible for making the decisions. Bringing the scientific findings and derived recommendations to the attention of the international climate change community and feeding them into relevant policy reports, such as the Fifth IPCC Assessment Report currently under preparation, are therefore of crucial importance.

This applies in particular to the opportunities that groundwater, being the world's largest available store of freshwater, can offer in terms of adaptation to the impacts of climate change and human activity. Controlled recharge and subsurface storage of water in aquifers, and recovery of this water in times when water is scarce, has proven to be a promising way to adapt to increasing variability of precipitation patterns and a higher frequency of extreme weather events, such as floods and droughts. Maintaining or augmenting recharge of aquifers can also alleviate the intrusion of seawater into coastal aquifers and combat land subsidence caused by falling hydraulic heads due to overexploitation of aquifers. On a smaller scale, water harvesting techniques might catch water during rainfall events in order to recharge an aquifer, thus impeding the quick runoff out of a catchment area. This is particularly important for people living in arid and semi-arid regions characterized by erratic rainfall and prolonged periods of drought, where every drop of water counts. While increasingly being applied in small and large schemes around the world, the opportunities that groundwater offers for climate change adaptation remain only marginally recognized and are not fully taken into consideration in decision-making processes.

Achieving (i) a better scientific understanding of the subject and (ii) increased recognition of the role of groundwater in the context of global changes, and climate change in particular, are closely linked:

- Sound knowledge is a sine qua non condition for communicating strong messages. These messages should highlight both the opportunities that groundwater can offer for climate change adaptation and the challenges and risks that groundwater resources are facing in a changing environment.
- Conversely, increased public and political recognition will lead to improved awareness, an increase of funding to advance scientific research and management, and broader application of schemes that include groundwater for adaptation to a changing climate and society.

This publication presents an overview of recent advances in knowledge related to the assessment and management of groundwater resources, giving special attention to the uncertainties related to climate change and variability. While providing a comprehensive description of hydrogeological characteristics of groundwater systems, the present volume also covers important aspects of legal and institutional contexts required for groundwater resources management as well as social and economic considerations.

It is hoped that the contents of this volume will contribute to an improved understanding of the impacts of climate change and human activity on groundwater resources and will provide useful guidance for policy makers and planners to include groundwater in climate change adaptation schemes and strategies.

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Reference

The IPCC Technical Paper on Climate Change and Water. https://www.ipcc.ch/pdf/technical-papers/ccw/frontmatter.pdf (2008)

Preface

Groundwater is a major source of water across much of the world and acts as a component of the global water cycle on the Earth. Groundwater has the capacity to balance large variations of precipitation and associated increased demands during water shortage, and may provide valuable alternative water sources when surface-water resources are close to the limits of sustainability. However, groundwater resources may be threatened by the uncertainty of climate change and increased water demand from human activities.

This book focuses on three major objectives: (1) to overview the current knowledge of groundwater resources and management in a changing climate and society, (2) to make adaptation, alternatives, and resilience the strategies for groundwater management in changing environments, and (3) to discuss new directions and initiatives of hydrological study, particularly of groundwater.

One of the groundwater resources assessment programs related to climate change is the GRAPHIC project (Groundwater Resources Assessment under the Pressures of Humanity and Climate Change), which was initiated by UNESCO's International Hydrological Programme (IHP) in 2004. The project provides a platform for communication and exchange of knowledge among groundwater and climate experts around the world. This book includes some of the research results from GRAPHIC projects.

The book also contains contributions from the intensive training course on groundwater resources, science, and management which accounted for a portion of Japan's contribution to UNESCO's IHP. The course was composed of a series of lectures, symposiums, and practice sessions led by experts in the field and laboratory, and of several technical field visits. The 20th IHP training course was organized by the Research Institute for Humanity and Nature (RIHN), Kyoto; Nagoya University; and Kyoto University.

We gratefully acknowledge the reviewers of book's chapters as well as UNESCO GRAPHIC and the 20th IHP training course program. In particular, we would like to acknowledge the following colleagues for their generous assistance: Prof. Hiroshi Uyeda, Nagoya University; Prof. Hirohiko Masunaga, Nagoya University; Prof. Kaoru Takara, Kyoto University; Prof. Tomoharu Hori, Kyoto University; Prof. Yasuhiro Takemon, Kyoto University; Prof. Takanori Nakano, RIHN; Prof. Junpei Kubota, RIHN; and Prof. Tsugihiro Watanabe, Kyoto University.

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Global Environmental Studies

The Global Environmental Studies series introduces the research undertaken at, or in association with, the Research Institute for Humanity and Nature (RIHN). Located in Kyoto, Japan, RIHN is a national institute conducting fixed-term, multidisciplinary, international research projects on pressing areas of environmental concern.

RIHN seeks to transcend the common divisions between the humanities and the social and natural sciences, and to develop synthetic and transformative descriptions of humanity in the midst of a dynamic, changeable nature. The works published in the series will reflect the full breadth of RIHN scholarship in this transdisciplinary field of global environmental studies.

Chapter 1
Groundwater Research and Management: New Directions and Re-invention

Mary P. Anderson

Abstract Groundwater hydrology is a relatively new scientific discipline, although the use of groundwater for water supply dates from the time when humans first noticed the existence of springs. Some research themes in groundwater hydrology, for example, well hydraulics, consistently receive strong interest while other themes are periodically revisited and reinvented. For example, recent interest in the hyporheic zone motivated research on the old topic of groundwater–surface water interaction. Improved instrumentation for measuring subsurface temperature motivated a return to the old topic of heat as a groundwater tracer. Groundwater sustainability is a re-invention of safe yield. Assessing hydrologic impacts of climate change involves analyzing temporal variability, which hydrologists have done for a long time.

In this chapter, the following four "big" issues are considered for the purpose of exploring challenges for groundwater management in the twenty-first century: sustainability, climate change, effects of agricultural land use, and water and energy. Addressing those four issues involves both policy and research challenges.

Keywords Climate change • Groundwater sustainability • Irrigated agriculture • Water and energy

1.1 Introduction

Groundwater hydrology grew out of hydraulic engineering with efforts to develop groundwater resources (Darcy 1856). Because the main method of accessing groundwater is by wells, it is not surprising that much research in groundwater

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Table 1.1 New issues derived from old concepts

Old concepts	Corresponding "new" issues
Safe yield	Sustainability ethics/stewardship Hydrogeoecology
Temporal variability	Climate change/regional hydrology
Agriculture	Agriculture consumptive use
	Groundwater contamination
Waste disposal/groundwater contamination	Water and energy
	Geologic repositories for high-level nuclear waste
	Carbon sequestration
	Hydraulic fracking

hydrology has been focused on the theory and application of well hydraulics. The early work of Dupuit (1863), Forchheimer (1886), and Thiem (1906) led to a major breakthrough by Theis (1935), who developed a method for the analysis of transient response of groundwater levels to pumping for the purpose of calculating hydraulic conductivity. New methods in well hydraulics include hydraulic tomography (Yeh and Lee 2007; Illman et al. 2008; Bohling and Butler 2010) and direct push techniques (Butler et al. 2007).

In the twenty-first century, attention is focused on the challenge of achieving groundwater sustainability (e.g., Gleeson et al. 2012), especially given concerns over global climate change and serious groundwater depletion in a number of basins worldwide (Narasimhan 2009). The goal of groundwater sustainability is to provide access to good-quality groundwater, but implicit in that goal is protection of groundwater resources from contamination and remediation of contaminated groundwater.

The U.S. National Academy of Engineering recently formulated 14 grand challenges for engineering (NAE 2008), including 3 related to groundwater hydrology: providing access to clean water, managing the nitrogen cycle (as related to agricultural land use), and capturing carbon dioxide. For the purposes of this chapter, those three challenges are recast as four "big" issues related to groundwater: sustainability, climate change, agricultural land use, and water and energy (including carbon sequestration).

These are not new issues (Table 1.1). Groundwater sustainability grew from the old concept of safe yield (Alley and Leake 2004) while climate change is essentially long-term temporal variability. Although hydrologists have long been concerned with the effects of agricultural land use, water and energy is a new topic. However, groundwater issues related to energy generally focus on the potential for groundwater contamination during energy development and subsurface disposal of associated waste. Groundwater contamination has been a management and research focus since at least the 1970s. In this chapter, each of these four "big" issues, as well as the associated policy and research challenges, are discussed.

1.2 Sustainability

Groundwater sustainability can be defined as the beneficial use of groundwater to support present and future generations while ensuring that unacceptable consequences do not occur. In a discussion of sustainable water development, it is important to remember that evapotranspiration consumes 60–66 % of the water that falls as precipitation, leaving only 34–40 % for other purposes (Narasimhan 2010). Globally, groundwater forms the largest available reservoir of freshwater on Earth, constituting 30.1 % of the freshwater supply whereas surface water constitutes 0.4 %. Currently, 69.5 % of freshwater is frozen in glaciers, snow, ice, and permafrost (Black and King 2009, p. 21) and hence is unavailable for water supply. Globally, 65 % of groundwater is used for drinking water, 20 % for irrigation and livestock, and 15 % for industry and mining (Zektser and Everett 2006).

The concept of safe yield was introduced by Lee (1915) to provide a way to determine the amount of groundwater that could be developed "safely." Safe yield became the guiding principle for groundwater management in the United States in the 1950s. The perennial safe yield was defined as the amount of groundwater that could be removed on an annual basis without causing undesirable effects. Safe yield was understood to be a single finite number that would define allowable groundwater withdrawals. The safe yield was often equated with the amount of recharge to the groundwater system in a given watershed, even though early researchers (Theis 1940) had pointed out that wells are never able to capture all the natural recharge. Furthermore, because groundwater development first removes water from storage in the cone of depression around the pumping well, there is a time lag, sometimes a very long time lag, before the effects of pumping reach the discharge area. Moreover, the safe yield is unbounded if it is acceptable to take significant amounts of water from storage in the aquifer or as induced recharge from surface water (Bredehoeft 2002).

Development of groundwater resources by pumping inevitably causes declines in groundwater levels and storage. Sustainable groundwater use implies that it is possible to reach a new steady state during groundwater development where declines in groundwater levels and storage stabilize. If a new steady state is reached, natural discharge from the basin will decrease unless a new source of water is introduced to the basin or water is returned after use. For example, treated sewage water could be injected into the groundwater system to return water to the aquifer. Given the large groundwater withdrawals in many basins worldwide and the continuing declines in water levels and groundwater storage, it is unfortunately the case that for some basins current demand can only be met by continual depletion of storage. Thus, it is generally recognized that groundwater pumping may not be sustainable (Kendy 2003; Zheng et al. 2010). Aquifers subjected to unsustainable pumping include the groundwater system underlying the North China Plain, the High Plains aquifer (USA), and the Midwest Cambrian-Ordovician aquifer (USA).

There are two problems inherent in the definition of sustainable groundwater use. Similar to safe yield, sustainability relies on defining "unacceptable consequences."

Someone must decide what is meant by unacceptable consequences in each specific situation. Pumping always causes declines in groundwater levels and storage, and a decrease in natural groundwater discharge is likely to occur in most basins. Groundwater development may also result in seawater intrusion, degradation in water quality and groundwater contamination, land surface subsidence, changes in vegetation, decrease or loss of streamflow, loss of springs, impairment and loss of wetlands, and loss of animal habitat. Deciding at what point in time the decline in water level, loss of storage, and associated impacts are "unacceptable" is a subjective policy decision.

Second, the definition of sustainability refers to "present and future generations" with the implicit understanding that we refer to generations of humans. However, ethics opens up the broader issue of treating all entities of the ecosystem, both animate and inanimate, as members of a community with equal rights to beneficial use of water. The notion that the "land" has rights, whereby land is understood to include all parts of the environment, was introduced by Aldo Leopold as the Land Ethic (Leopold 1949; also see Anderson 2007; Liu et al. 2010). In the broadest extension of the Land Ethic, inanimate entities such as rocks and groundwater have rights to remain in an undisturbed state in at least some parts of the world (Nash 1977).

Acceptance of a true groundwater ethic whereby groundwater, springs, rivers, lakes, rocks, soil, plants, and animals are recognized to have rights is a long way off (Liu et al. 2010). However, ideas of stewardship and living in harmony with the environment (Fitch 2009) are steps toward acceptance of the Land Ethic, in its broadest interpretation. Stewardship, in general, implies care of something that belongs to someone else. Stewardship of the Earth embodies a responsibility to preserve and protect our natural resources for the benefit of future generations. Thus stewardship, by individuals, communities, and government, is a means toward sustainability. The European Water Framework Directive, for example, is a governmental initiative that calls for enactment of regulations that embody principles of sustainability and stewardship.

1.3 Climate Change

The Intergovernmental Panel on Climate Change (IPCC 2001) defines climate change as "a statistically significant variation in the mean state of the climate or its variability, persisting for an extended period (typically decades or longer)." Under this definition, climate change "may be caused by natural internal processes or external forcings or by persistent anthropogenic changes in the composition of the atmosphere or land use" (IPCC 2001). Following this definition, climate change essentially is long-term variability that is expected to occur globally on scales and intensities unprecedented in human experience.

The evidence for global climate change has been presented in numerous publications: a recent example is the paper by Chapman and Davis (2010).

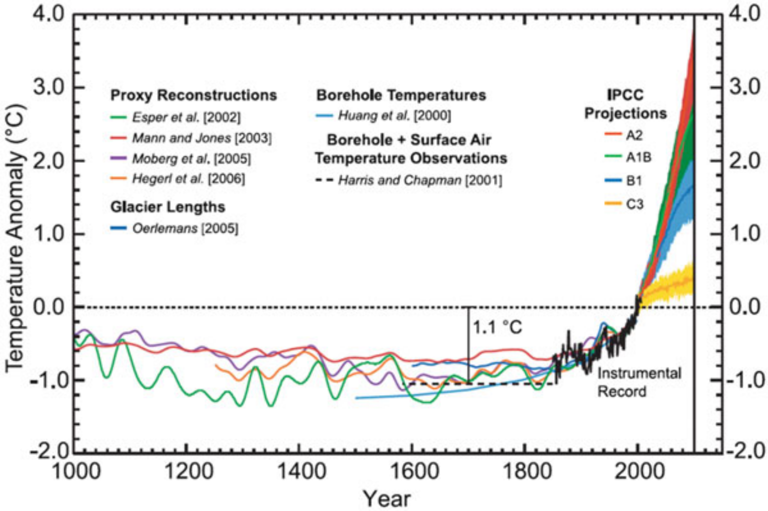

Fig. 1.1 Climate variability as measured by average global surface temperature based on reconstructions from proxies such as tree rings and corals, glacier lengths, and temperatures in boreholes. Measured temperatures in the nineteenth and twentieth centuries are also shown as are temperatures predicted for the twenty-first century based on global climate models (A2, A1B, B1, and C3) as reported by the Intergovernmental Panel on Climate Change (IPCC). (From Chapman and Davis 2010)

Nevertheless, skeptics remain. Indeed, review of historical trends in climate do not always support climate change. For example, analyses of the historical record of precipitation in Wisconsin, USA, indicated long-term variation in the magnitude and frequency of large daily rainfall events but found no evidence to support changes caused by global climate change (WICCI Stormwater Working Group 2010). Loáiciga (2009) concluded from a review of temperature and carbon dioxide trends in the geologic record up to the present time that these variables "have exhibited anything but drastic change over geologic time." Some scientists continue to challenge that global climate change owing to anthropogenic activities is occurring, despite evidence from the instrumental record of an accelerated increase in average global temperature (Fig. 1.1). Moreover, global climate models predict a rapid increase in the average global temperature within the next 50 years (Fig. 1.1) even under the unrealistic scenario that carbon dioxide levels are held to concentrations measured in the year 2000. However, some scientists have pointed out that predictions of global climate models are highly sensitive to even small errors in initial conditions (Loáiciga 2009; Collins 2002). Nevertheless, given that temporal fluctuations in climate are inevitable, it seems prudent to plan for global climate change.

The hydrologic effects of climate change include droughts and floods (including groundwater flooding caused by rising water tables) and seawater intrusion into estuaries and coastal aquifers. Droughts compound declines in groundwater levels caused by pumping. The response to climate change involves policy options under the broad categories of adaptation and mitigation. Adaptation strategies are discussed below under Policy Challenges. Mitigation involves efforts to reduce emissions of greenhouse gases, mainly carbon dioxide, to the atmosphere. The main mitigation strategy relevant to groundwater is carbon capture and sequestration (CCS), which is discussed below under Water and Energy and under the section on research challenges for groundwater contamination.

1.4 Agricultural Land Use

Cultivation of land for agriculture is driven by the desire for food security, but one-fifth of the world's cropland cannot support agriculture without irrigation (Black and King 2009, p. 62). Irrigation accounts for about 70 % of global freshwater use; about 38 % of irrigated land uses groundwater (Siebert et al. 2010). In many countries more than 50 % of groundwater pumped is used for irrigated agriculture (Table 1.2). Moreover, globally the land area under irrigation is increasing (Black and King 2009). Irrigation accounts for around 90 % of global consumptive water use (Siebert et al. 2010), such that water applied to crops is lost by evapotranspiration to the atmosphere. Large depletions in groundwater have occurred in areas where groundwater is exploited for irrigated agriculture (e.g., the High Plains aquifer, USA, and the North China Plain). Thus, agricultural land use is closely tied to concerns about groundwater sustainability.

Furthermore, application of fertilizers (nitrogen and phosphorus) and pesticides may result in widespread groundwater contamination beneath agricultural land. Salinization of the soil is also common because crops remove water and leave behind minerals that accumulate in the soil.

1.5 Water and Energy

Development of energy resources potentially causes groundwater contamination as a result of methods used to extract the resource and from disposal of associated wastes. The processes that control the transport of contaminants in groundwater

Table 1.2 Percent of groundwater used for irrigation in selected countries

India	89
Spain	80
Argentina	70
United States	68
Australia	67
Greece	58
China	54
Japan	23

Source: Zektser and Everett (2006)

have been an area of intense research since the 1970s (Apgar and Langmuir 1971; Baedecker and Back 1979; Anderson 1979). Long-term field investigations at dedicated research sites have increased our understanding of both physical and chemical transport processes (Anderson and McCray 2011). Sophisticated transport models have been tested and applied to many field situations. Yet, our current transport models have the same problems as the early models in use in the 1970s. Specifically, these include (Anderson 1979): "… problems in acquiring detailed field data, theoretically defining dispersivity, solving the dispersion equation numerically, and incorporating chemical reactions into the modeling scheme."

Challenges related to energy development in the twenty-first century include design of geologic repositories for permanent storage of high-level nuclear waste, design of facilities for carbon sequestration in the subsurface, and hydraulic fracking for retrieval of gas and oil. Each of these challenges involves potential risks to groundwater.

Despite much scientific investigation, no country has yet established a repository for permanent storage of high-level nuclear waste. The preference is for subsurface disposal in a geologic repository designed to minimize leakage to groundwater. The hydrogeological setting is a key component of the design. Several countries are focusing on disposal in granite. The approval of a geologic repository involves both science and engineering as well as political and societal issues. For example, the intensively studied site in the USA in a salt deposit at Yucca Mountain, Nevada, was recently taken out of consideration for political reasons.

One mitigation strategy for slowing global warming is to capture carbon emissions from the burning of fossil fuels and store them in the subsurface, a process known as CCS. Pacala and Socolow (2004) introduced the concept of the carbon wedge as a way to quantify the mitigation effort. In their conceptual model, we must eliminate emissions equivalent to seven carbon wedges to stabilize carbon emissions at 2004 rates. An operating CCS site at Sleipner, Norway, currently captures and stores 0.3 million tons of carbon/year (Pacala and Socolow 2004). It is sobering, however, to realize that approximately 3,500 such facilities would be required to account for just one carbon wedge. Widespread CCS will require careful hydrogeological site investigations and will present challenges in modeling because the transport processes involving carbon dioxide sequestration are complex (Zhou et al. 2010; Celia and Nordbotten 2009). Migration of carbon dioxide out of the host rock into neighboring aquifers is a concern, as is possible return of carbon dioxide to the surface.

In hydraulic fracking, water and chemicals are introduced to a gas/oil reservoir to create fractures to aid in recovery of the resource (shale gas or shale oil). Typically, the site of injection is several thousand meters below the surface. The process has been facilitated by relatively recent improvements in directional horizontal drilling methods. Hydraulic fracking makes possible the development of low permeability shales and has led to an oil and gas boom in the United States (Walsh 2011). However, the process of injecting fluids into the subsurface during development of the reservoir, as well as the disposal of wastes by injection back into the subsurface, has caused a series of small earthquakes in at least two locations. Furthermore, escape of the injected chemicals into the surrounding rocks has led to groundwater contamination of shallow aquifers.

1.6 Policy Challenges

Given that groundwater resources are already being depleted in unsustainable amounts in many places in the world and given that the world's population is projected to increase from 7 billion in 2011 to 9.4 billion in 2050, it is imperative to devise groundwater management policies that will ensure adequate water supplies under the new conditions anticipated as a result of climate change. The goal of both sustainability and planning for climate change is to ensure adequate supplies of good quality groundwater. Hence, sustainability and climate change are really two sides of one issue. Furthermore, agricultural land use is closely linked to groundwater sustainability because much of the water used for irrigated agriculture is groundwater.

1.6.1 Sustainability

Management strategies that contribute toward sustainability include conjunctive use of groundwater and surface water, including the use of aquifer storage recovery (ASR), whereby excess surface water is injected into the subsurface for storage and retrieval during times when surface water is scarce. Transfer of surface water to conserve groundwater resources is also an option that is spectacularly being executed via the South to North water transfer project in China (Zheng et al. 2010). Another option is water re-use whereby advanced treatment of sewage produces water of drinking water quality. Overcoming public perception that re-used water is still "sewage" is a major sociological challenge. Injecting the treated water into the subsurface provides additional treatment before pumping it back out for water supply through ASR, a process that helps alleviate the public's concerns over water re-use. Desalinization is another strategy for producing "new" water. All the aforementioned strategies should be utilized, as appropriate, under adaptive management plans, whereby it is expected that the plan will evolve to adapt to new environmental, technological, and sociological conditions as they arise.

Sustainable use of groundwater implies that good-quality groundwater is available. Hence, there is a need for regulations to protect groundwater from contamination by agricultural chemicals (fertilizers and pesticides) as well as chemicals from energy development and other types of waste (e.g., municipal and industrial waste). Finally, another dimension to sustainability is groundwater ethics. Ethics as currently practiced is almost always focused on human needs, but principles of stewardship of the Earth and harmony with the environment are gaining acceptance.

We should strive for acceptance of a true groundwater ethic that invokes the Land Ethic, whereby all animate and inanimate entities are treated as members of a community with rights to beneficial use of groundwater.

1.6.2 Climate Change

Policy options to address climate change fall under the two broad categories of adaptation and mitigation. The strategies already discussed for sustainability can be considered adaptation strategies for climate change. Another adaptation strategy is the migration of populations from water-poor to water-rich areas. Mitigation strategies are aimed at reducing the emissions of carbon dioxide to the atmosphere. The mitigation strategy most relevant to groundwater is CCS, as discussed in the section on Water and Energy.

1.6.3 Agricultural Land Use

Irrigated agriculture is a large and consumptive use of water worldwide. Hence, policy options should be aimed at reducing the area of irrigated land (Kendy 2003) and increasing water use efficiency to reduce the loss of water from the watershed by evapotranspiration and degradation of soil by salinization. Furthermore, regulations to control the application of agricultural pesticides and fertilizers (nitrogen and phosphorus) are important to minimize groundwater contamination under agricultural land.

1.6.4 Water and Energy

Policy and regulations control the use of engineering options available for development of energy resources. The political and societal situation is especially important in the approval of a site for a geologic repository for high-level nuclear waste. To date, no country has managed to overcome the engineering challenges and political/societal concerns associated with opening such a repository.

Similarly, policy will play a role in whether a country decides to invest in CCS. There are only a few operating CCS facilities worldwide, pending ongoing research efforts to define conditions necessary for environmental protection. Given the large number of such facilities required to make a difference, it is likely that some countries (e.g., China) will be better able to muster the necessary political willpower than other countries (e.g., the United States).

Environmentally safe utilization of hydraulic fracking also will require regulatory oversight. Concerns related to earthquake generation and groundwater contamination have stalled the widespread use of hydraulic fracking in the United States pending review by the federal regulatory agency.

1.7 Research Challenges

Just as sustainability and climate change have the same management goals, they also share the same research challenges. Research challenges related to transport of agricultural contaminants in the subsurface and those associated with energy development are research challenges for contaminant transport.

1.7.1 Sustainability and Climate Change

Long-term data are essential to develop and refine water budgets needed for planning to meet the challenges of sustainable groundwater development under climate change. Hence, long-term monitoring networks that, at a minimum, measure precipitation, groundwater levels, and streamflow are needed. Monitoring lake levels and evaporation from the lake surface, springflow, and new and improved sensors for monitoring hydrologic and related variables such as temperature are also needed. New technologies can foster advances in research aimed at understanding hydrologic processes. For example, new sensors for measuring temperature led to renewed interest in using heat as a groundwater tracer (Anderson 2005) to study hydrologic processes such as groundwater discharge to streams. New attention to ethics and stewardship has led to efforts to protect the ecosystem from "unacceptable consequences" caused by groundwater development, and this has fostered a new discipline known as hydrogeoecology (Hancock et al. 2009; Lewis 2011).

Understanding and quantifying hydrologic processes affected by climate change require research focused on the linkages between reservoirs in the hydrologic cycle and on feedback mechanisms. Feedback between groundwater and surface water has been a long-standing focus area for groundwater hydrologists, but new issues continue to emerge, as exemplified by recent work on the hyporheic zone, which is the interface between surface water in a stream and groundwater. However, relatively little is known about possible feedback mechanisms between groundwater and the atmosphere, which could affect climate. For example, DeAngelis et al. (2010) suggested that irrigation using groundwater from the Ogallala Aquifer (USA) caused a regional increase in precipitation owing to consumptive use of groundwater and loss to the atmosphere via evapotranspiration. Furthermore, groundwater brought to the surface via pumping releases carbon dioxide (Wood 2009). Another example is that the large-scale transfer of groundwater from land to ocean, which is caused by pumping from large groundwater basins worldwide, is estimated to be responsible for 5–10 % of observed sea level rise (Konikow 2011; also see Konikow and Kendy 2005 and Wada et al. 2010).

Addressing global climate change requires groundwater hydrologists to address problems at regional, continental, and global scales rather than the site scale typical of many traditional groundwater problems. For example, Fan et al. (2007) compiled a water table map for the North American continent. Regional hydrogeology is an emerging new discipline in groundwater hydrology.

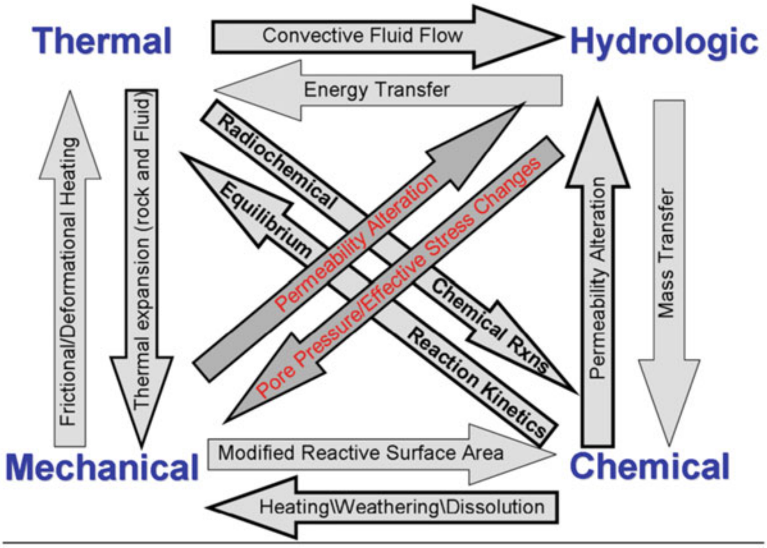

Fig. 1.2 Coupled processes that should be considered when modeling transport of contaminants in the subsurface. (From Yow and Hunt 2002)

1.7.2 Contaminant Transport

Although much has been learned about the processes governing the transport and transformation of chemicals in the subsurface (including agricultural chemicals), many questions remain. A consensus to emerge from the large body of research on this topic is that geologic heterogeneity affects the physical transport of solutes as well as their chemical transformation in the subsurface. A major challenge, therefore, is to characterize the subsurface in sufficient detail to identify the location, geometry, and hydraulic properties of geologic heterogeneities. New tools to help characterize the subsurface are needed; geophysical methods are one of the few tools capable of "seeing" between boreholes in the subsurface. Recent improvements in the application of geophysical methods to groundwater investigations have fostered the new field of hydrogeophysics (Rubin and Hubbard 2005; Knight et al. 2010).

To capture the detailed heterogeneity that is important to transport, some researchers advocate characterizing the subsurface at the meso-scale (from tens of centimeters to meters) and modeling transport processes at the pore scale (Scheibe 2009). Methods for upscaling from the pore scale to the field scale will be critical to this work. Also needed are new understanding and new predictive tools to quantify the complex coupled hydraulic, chemical, thermal, and mechanical processes that govern the transport and transformation of chemicals in the subsurface (Fig. 1.2).

1.8 Summary and Conclusions

Groundwater constitutes the largest available freshwater reservoir on Earth. Demands for groundwater have caused serious depletion of many aquifers around the world as a result of pumping. Furthermore, the desire for food security drives expansion of agricultural land use and large consumptive uses of groundwater for irrigation (Table 1.2). Groundwater is also at risk from contamination by agricultural chemicals and municipal and industrial wastes, as well as a result of energy development.

Challenges in groundwater management in the next 50 years can be lumped into four “big” issues: sustainability, agricultural land use, climate change, and water and energy. None of these are new issues, as sustainability grew out of the concept of safe yield and climate change can be viewed as long-term temporal variability. However, sustainability has new dimensions that include the realization that, under some circumstances, groundwater pumping is unsustainable. Furthermore, sustainability encompasses ethics including concepts such as stewardship and living in harmony with the environment, as well as principles embodied in the Land Ethic. Moreover, sustainability must be attained under conditions brought about by climate change. Global climate change is anticipated to cause changes in average temperatures unprecedented in human experience that will drive changes in average precipitation/evaporation, and the frequency and intensity of floods and droughts, as well as seawater intrusion into coastal aquifers.

Sustainability implies access to good-quality water. Hence, agricultural land use and development of energy resources, both of which can potentially cause groundwater contamination, are tied to issues of sustainability. Furthermore, agricultural pesticides and fertilizers potentially contaminate groundwater, and the soil is degraded by salinization. New issues related to groundwater contamination will emerge as we continue to work toward the design and eventual implementation of geologic repositories for the permanent storage of high-level nuclear waste, investigate the feasibility of carbon sequestration in an array of hydrogeological settings, and find ways to use hydraulic fracking without adverse effects.

Policy options for sustainable use include conjunctive use of groundwater and surface water, water re-use, especially with ASR, water transfer, reducing irrigated agricultural land and increasing water use efficiency, as well as fostering concepts of ethics, stewardship of the Earth, and living in harmony with the environment.

To achieve sustainability and plan for climate change, we need to establish and maintain long-term monitoring networks and to develop new and improved sensors for hydrologic monitoring. Research is needed to understand and quantify linkages among reservoirs in the hydrologic cycle at regional, local, and microscopic scales. Ethics drives efforts to understand the linkages between hydrology and ecosystems (eocohydrology). New and improved tools (especially geophysical tools) are needed to help characterize geological heterogeneity to quantify the transport and transformation of chemicals in the subsurface. New and improved modeling approaches are needed to quantify complex, coupled hydraulic, thermal, chemical, and mechanical subsurface transport processes, especially as related to scaling issues. These research directions point to the need for increased emphasis on interdisciplinary research focused on groundwater. Schwartz (2012) recently noted that: “Today, there is less emphasis on pure research in small teams and more emphasis on larger and more complex problems of global importance undertaken by multidisciplinary teams. The challenge of emerging new problems is unprecedented in the history of hydrogeological sciences.”

As a growing population of humans faces the challenges of water sustainability in the twenty-first century, it is inevitable that the world will need access to more good-quality water and that issues related to groundwater will advance in importance.

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Chapter 2
Groundwater as a Key of Adaptation to Climate Change

Makoto Taniguchi

Abstract Sustainable groundwater use is discussed in terms of the balance between risks and services, including adaptation to climate change. Groundwater, with its longer residence time of water circulation, can be an alternative water resource and environment under climate change. Assessments of groundwater services and benefit as well as risk are important for sustainable groundwater use under climate change conditions. Groundwater, which is one of the keys of adaptation to climate change, should be treated as a common resource and environment beyond the tragedy of the commons and the dilemma of boundaries.

Keywords Climate change • Commons • Public and private waters • Resource and circulation • Risk and service • Sustainable groundwater use

2.1 Introduction

Global environmental problems are the consequence of the quandary of striking a balance between humans and nature amidst dynamic natural and societal change so as to build a more functional society. In the case of water problems, it may also be said that the negative aspects of floods and disasters also rest upon balancing them against the benefits and services derived from water resources and water circulation.

Figure 2.1 shows the relationships of various components within three models related to global environmental problems (Moss et al. 2010). Atmospheric, oceanic, snow–ice, and continental carbon circulation are the subjects of the climate model, and energy, economics, health, agriculture, and forestry are central to the integrated model. Rise in sea level, the ecosystem, human dwellings, infrastructure, and other aspects are focused upon in the influence–fragility and risk assessment model.

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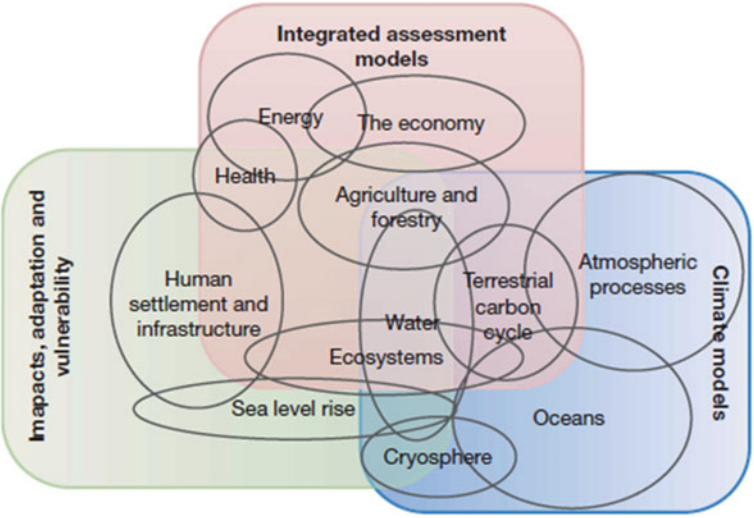

Fig. 2.1 Framework of the models on global environmental problems (Moss et al. 2010)

"Water" is central in all three models: thus, when global environmental problems are broadly separated into the three categories of tangible things, society, and living beings (including humans), the "problem of water" is central to all their relationships.

Among them, the water problems considered up to this point, namely, there being either "too much water" or "too little water," were based on the unbalanced distribution of people and water. However, water problems are not limited to these concerns and are problems spanning space and time. Figure 2.2 depicts the spatial positional relationships (from close to distant) of people and water on the horizontal axis and the circulation time/movement speed of water (from fast to slow) on the vertical axis. From early times in the past, fast-flowing water that was nearby, such as rainwater and river water, and that from ponds and shallow groundwater was used; however, the increase in demand for water together with the need for supplemental water resources at a distance led to the building of dams and waterworks facilities and the transport of water from these structures. Moreover, recently "far-away water" that crosses national borders and watershed areas has come into use, as can be seen from the use of bottled water. Meanwhile, slower groundwater from greater depths is currently being used even when the same water sources are nearby. The virtual water in imports of agricultural and livestock products derived from the use of deep aquifers such as America's Ogallala (High Plains) Aquifer and groundwater under the North China Plain illustrates the current situation wherein water that is even slower and more distant is being used. "Far water" presents the problem of difficulties in administration that extends beyond human awareness, and "slower water" is problematic because of the time needed for it to recover. Both these are problems that creep up unawares and will have a great influence on the future.

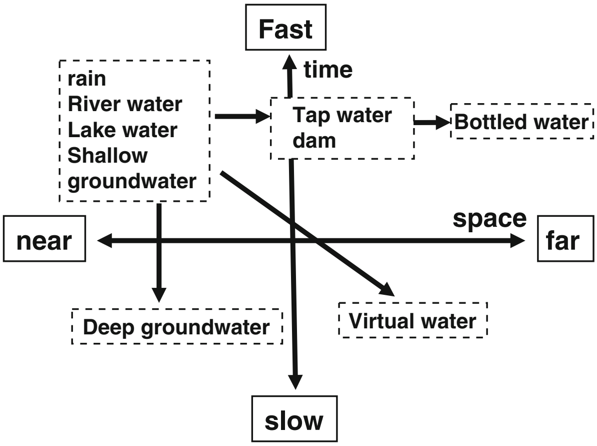

Fig. 2.2 Water uses in time and space

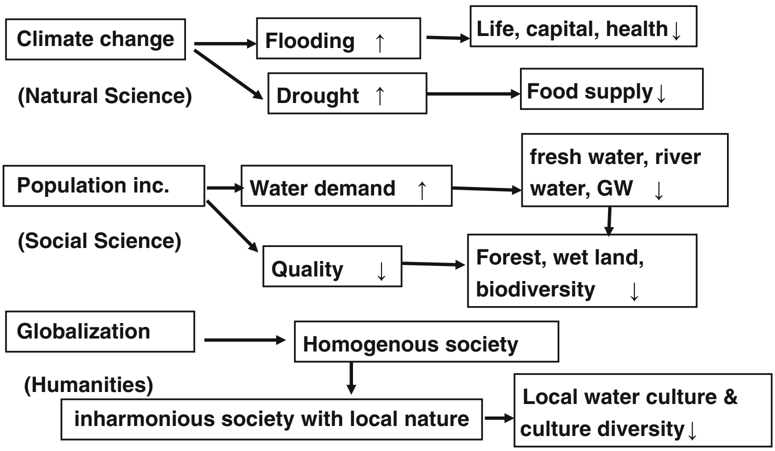

Fig. 2.3 Causes of water problems. *GW*, groundwater

Three factors involved in creating the water problem that is central within the global environmental problem can be posited in a broad sense (Fig. 2.3). One of these is the natural science factor of climate change. Climate change, which includes global warming, changes the patterns of rainfall and greatly increases the risks of floods as well as the frequency and severity of droughts. The former is linked to loss of life, health, and property whereas the latter is a serious global environmental problem that puts pressure on the supply and demand for food. The second is the societal factor of the increase and concentration of the human population. The escalation in the withdrawal of water resources stemming from increased water demand causes a reduction in stored groundwater and the blockage of rivers as well as a reduction of freshwater areas, which in turn results in worsening of the environments of forests, farmlands, marshes, and coastal water areas coupled with declining biodiversity. In addition, the excessive material load created by increased human activities invites the worsening of water quality and diminishes the qualitative factor in each of the freshwater resources just mentioned. The third factor is the globalization of society, which is a human cause. It has been pointed out that the uniform concept of values spreading across the globe forms societies that are not in harmony with a particular region, which may lead to the loss of the region's water culture and its cultural diversity. The current state of affairs views these three factors respectively as interrelating with each other to cause various global environmental problems.

2.2 Difference Between Groundwater and Surface Water in Relationship to Adaptation to Climate Change

When comparing surface water and groundwater in terms of water resources used by humans, their greatest hydrological differences lie in their residence times and replacement times. Table 2.1 shows the respective storage volumes, transport volumes (flux), and residence times within global water circulation. The residence (or replacement) time can be derived from the storage volume and flux (groundwater recharge rate); however, the differences between arid and humid regions as well as disparities in residence times arising from depth in the case of groundwater or other factors will create great differences among each factor (Table 2.2).

Looking at the amount available from the standpoint of storage volume, it appears that groundwater provides the largest source of usable freshwater resources; however, when considering it from the aspect of usage or flow in terms of flux, it becomes evident that the amount of river water overwhelmingly exceeds that of groundwater. About 435 mm of the 670 mm of rain that annually falls on land evaporates and the remaining 235 mm runs off as surface water. This 235 mm becomes a part of rivers and underground water and flows into the sea. The worldwide average groundwater recharge volume has been estimated to be 137 mm/year (Fig. 2.4; Doll et al. 2002). In the case of Japan, about 600–700 mm of an annual precipitation volume of 1,400 mm, is returned to the atmosphere through evapotranspiration and about 300 mm of the remaining 700–800 mm goes to groundwater recharge (Yamamoto 1983). The value

Table 2.1 Residence time of various waters

Seawater	2,500 years
Snow and ice	1,600–9,700 years
Permaforest	10,000 years
Groundwater	900–1,400 years
Soil water	1 year
Lake water	17 years
Wetland	5 years
River water	17 days
Vapor	8 days

Table 2.2 Causality of groundwater problem; D: Driving force, P: Pressure, S: State, I: Impact

D	P	S	I
Population increase	Increase in water demand	Increase in groundwater pumping	Decrease in groundwater storage
Climate change	Change in precipitation pattern	Decrease in groundwater recharge	
Urbanization	Land cover change		
Population increase in coastal zone	Increase in water demand	Saltwater intrusion	Groundwater salinization
Global warming	Sea level rise	Saltwater intrusion	
Increase in population	Increase in material loads	Improper water management	Groundwater contamination
Urbanization	Heat island	Subsurface warming	Effects on ecosystem?
Globalization	Capitalism	Water transfer with more energy consumption	Unsustainable groundwater uses

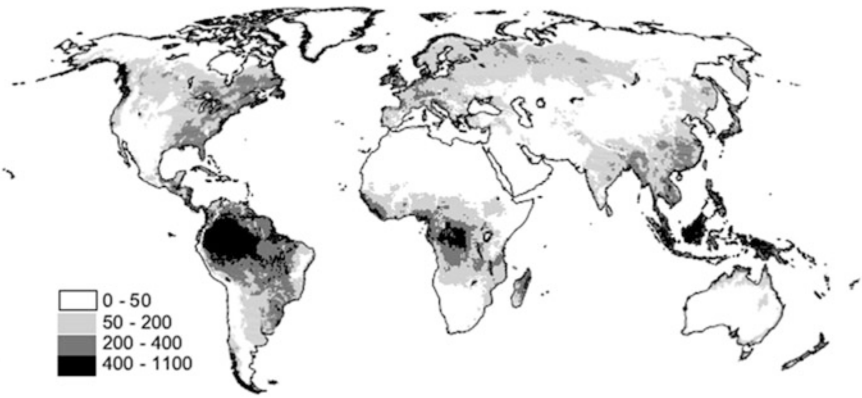

Fig. 2.4 Estimated groundwater recharge rate (mm/year) (Doll et al. 2002)

of the groundwater recharge rate (R) in relationship to the evapotranspiration (E) subtracted from the precipitation (P) indicates that $R/(P-E)$ in Japan has a smaller value than the worldwide average.

It has been pointed out that the change in precipitation patterns accompanying climate change bears the possibility of reducing groundwater recharge rates through an increase in surface runoff volume. In contrast, it has also been noted that there are examples where climate changes such as global warming have been related to the change in water resources.

Figure 2.5 shows the interannual changes in the number of days of precipitation for Taiwan (Taniguchi 2011). There has been little change in annual precipitation volumes for the past 100 years; however, a reduction in the days of precipitation throughout Taiwan is evident. It can be said that the rainfall patterns

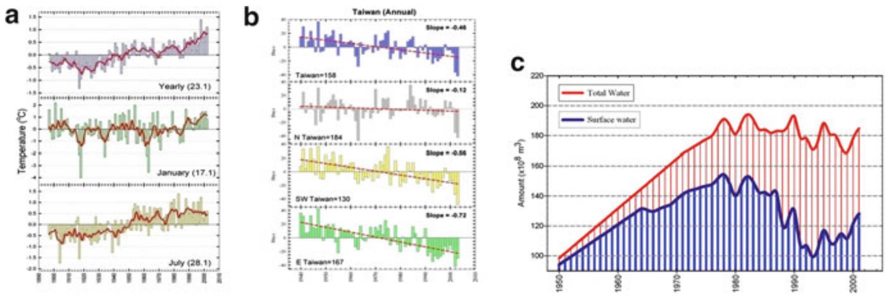

Fig. 2.5 Changes in water resources in Taiwan. (**a**) Increase in air temperature caused by global warming. (**b**) Decrease in precipitation date. (**c**) Change of water resources from surface water to groundwater

are drawing closer to those of a tropical rainfall pattern, with greater differences in the days with no rainfall and those with heavy rainfall. The reliability of water usage wherein surface water is stored in dams for use has been slightly shaken in conjunction with this, and there are indications of a trend toward the use of groundwater.

The case of Taiwan is an example of the water resource being switched as a consequence of climate change; however, the differences of residence time between groundwater and surface water were strongly related in this instance. Greater changes in the precipitation that makes up the base flow of the river water flow also mean that there will be larger changes in the short retention times for surface water. The magnitude of these changes in precipitation volume makes water administration even more difficult and increases the role of the relatively long groundwater retention time in its capacity as a natural low-pass filter.

2.3 Groundwater as a Natural Resource

When groundwater is viewed as a natural resource, the problem of where water is stored (retained) and who owns it becomes very important. Figure 2.6 shows a structural outline of the water problem in Saijo City in Ehime Prefecture. The problem is that surplus water from a dam constructed for the specific purpose of water for industrial use is planned to be transported (but this is not yet implemented) beyond the boundaries of the watershed. The Kamo River flowing through Saijo City becomes a raised riverbed and its river water recharges the groundwater. In Saijo, the groundwater derived from water recharged by the Kamo River creates an artesian well, the so-called “Uchinuki.” These “Uchinuki” wells have been used since ancient times. Although the waterworks facilities is considerably less than the national average, it may be said that this has been caused by the use of groundwater

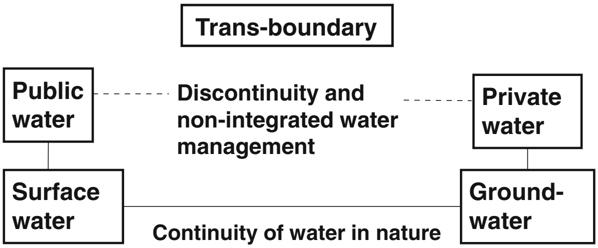

Fig. 2.6 Driving force, pressure, state, and impacts on groundwater problems

at every home. It is clear that a change in river discharge from the dam into the Kamo River would not only result in changes to the recharge volume for the groundwater, but that the change in groundwater flow volume on the coast would also alter the location the saltwater boundary, leading to saltwater intrusion. Two problems are involved: whether to use the water as surface water or as groundwater and the transboundary water predicament. The former problem includes the incommensurate dilemma wherein, under the Japanese system of water administration, surface water is handled as "public water," which means it is managed by public government, etc., whereas groundwater is associated with the property and considered to be "private water." This is a problem stemming from the scientifically contiguity of surface water and groundwater and its disconnectedness under the social system.

On the one hand, in the latter transboundary water problem, there is the public and private water issue, that is, who owns the water and how to overcome Hardin's "Tragedy of the Commons" (Hardin 1968), all concluding in the dilemma of how to administer groundwater as a moving common resource, capital, and asset.

Furthermore, the transboundary water problem leads to the "Dilemma of the Boundaries" when it is generalized (Taniguchi and Shiraiwa 2012). Borders such as national borders and regional administrative boundaries are necessary bounds for administration as they serve as the primary units of governments/administrations and their existence ensures cultural unity and clear administration. On the other hand, clear boundaries also mean that contiguous natural phenomena become disjointed for administration, leading to various inconveniences. As examples, a regional water administration system that differs with the watershed, which is a hydrological unit, the discontinuity of a system for continents and oceans, and differences in administrative water systems (the public water–private water problem) for the aboveground (surface water) and underground (groundwater), all indicate the various problem points inherent in boundaries. Although the importance of various sectors and the cooperation and coordination between interested parties has been pointed out in integrated water administration, an effective method for the rectification of differences in boundaries has yet to be found.

2.4 Services and Benefits of Circulating Groundwater

The groundwater in arid and semiarid regions may be somewhat considered only as a "resource" in its value and assessment; however, in humid regions, "circulation," which is another important aspect of groundwater, also comes into focus in value and assessment. In particular, it is necessary to weigh the risk assessments for floods, droughts, and other disasters with regard to the service provided by "circulation" and to evaluate them equally.

The millennium assessment for ecosystem services (Millennium Ecosystem Assessment 2005) shows the 16 services forming the basic services span a wide range and that it would be difficult to assess each individually in most cases. These services can be separated as "water" and classifying the services into those that do not mediate living organisms, or in other words, as things in an "environment system service"; and "ecosystem service," which mediates living organisms to a limited extent; and a "human service" in which humans are involved through city functions and so on. The "river maintenance water" used in the case of river water is close to the concept of environmental flow; however, its usage is diverse and dependent on which sections are to be included in the service. In the case of groundwater, the maintenance of the ecosystem at the stage where it reaches the surface as a spring, in river water, and in the riparian zone border (wetland) has drawn much attention.

The service of groundwater toward human society in its interim stage as it flows by is mainly as a water resource service; however, a part of it bears limited groundwater service assessment in its storage function use with the cool temperature of water in wells and in its emergency use for disaster control.

An example of the ecosystem service provided by groundwater discharged in coastal regions may be seen in habitats such as those of the *Crassostrea nippona* oyster in the shoreline area of the Chokai Mountain foothills. In Yuza Town at the Chokai Mountain foothills near the border between Akita and Yamagata Prefectures, *C. nippona* oysters, which live in a brackish water environment, can be found even in bays without river outflow, indicating the relationship between groundwater flow and the oysters (Fig. 2.7). The relationship can be seen in the annual precipitation in the watershed water balance and in the annual production of *C. nippona* oysters, which asserts the importance of groundwater as the freshwater component in the composition of the brackish water. In addition, isotope ratios analogous to those in the groundwater composition have been verified in the strontium isotope ratios that make up the *C. nippona* oyster shells, thereby providing evidence indicative of the involvement of the groundwater in the oyster habitat. This observation indicates that it is not only rivers which connect the land and the sea but that groundwater also exists as a mediator. In precipitous terrain such as that in Japan, it can be said that the volcanic geologic conditions with their good water permeability and the hydrometeorological conditions of plentiful precipitation maintain the shoreline habitat in relationship to the land and ocean via groundwater.

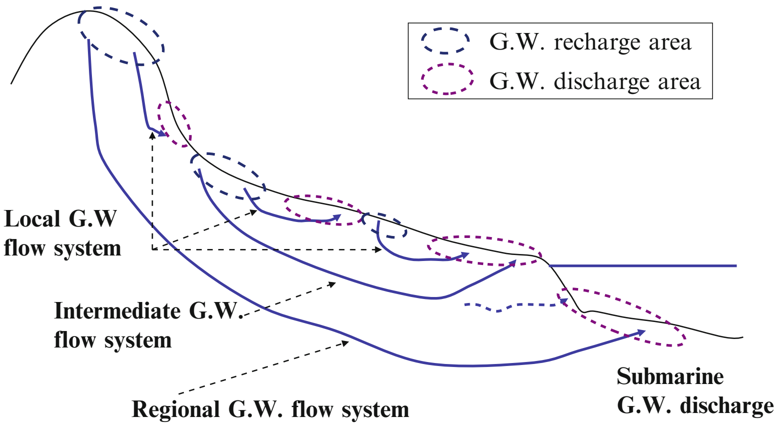

Fig. 2.7 Groundwater flow systems in humid regions

In this way, the assessment of the services provided by groundwater in maintaining the coastal habitat and other services, and the "service assessment" as an antithesis of "risk assessment" for disasters and such as they relate to groundwater, will continue to be important into the future.

2.5 Groundwater in Humid Regions

In humid regions such as Japan, the evaluation of not only the role played by groundwater as a "resource" but its assessment from the aspect of "circulation" and the "services" it provides is an important point. The key is to understand the intrinsic "volume permitted by nature (natural resource capacity)" in terms of the groundwater storage and recharge rates for groundwater while attempting to increase the capabilities of each respective regional society to their maximum using the latest technology, traditional wisdom, and the knowledge of international society such as late-comer's benefit.

Groundwater has the salient feature of having a longer residence time compared to surface water, so it functions as a "key" with respect to climate change and a changing society. Groundwater must be grasped in terms of its diverse problems with admissibility, adaptation, capability, and capacity, and proposals for a consortium to increase governance capabilities through legal systems and regulations and such have begun to be put forth (Taniguchi 2011). It will be a structure for discussions by the differing "sectors and stakeholders" who face common similar problems and share knowledge about them on the differences in policies stemming from the differences in the "volume permitted by nature" and the "latecomer's benefits" that differ in each of their countries, regions, and cities. On that occasion three types of components will need to be in place: (1) monitoring/assessment method, (2) modeling/visualization, and (3) policy planning, and it will be important that all these aspects be linked organically.

The subterranean environmental problems that recur in Asian cities are caused by a lack of understanding of the tolerances of the region and by exceeding them. An assessment method for the “volume permitted by nature” (capacity) of the underground environment for groundwater retention volume/groundwater recharge volume, etc., is close to being established. Meanwhile, consortiums and such should be used for education on the “latecomer’s benefits,” which are a part of the knowledge of international society, and on regional traditional wisdom as they relate to the capabilities of people and society.

Changes in water, materials, and the heat environment in the underground caused by human activities extends from depths ranging from a hundred to several hundreds of meters, and it is clear that groundwater circulation speed has increased more than tenfold in the past 100 years (Taniguchi 2011). In addition, it is also evident that the subterranean heat storage resulting from the heat island phenomenon accompanying urbanization has reached between twofold and sixfold that of the underground heat storage caused by worldwide global warming. Because direct use of the underground environment and human activities above ground influence the underground in this manner, it remains to be seen how far the underground environment can be used as an alternative environment for the surface. With regard to water volume, through the experience of the “Tragedy of the Commons” known as land subsidence, and at the current point of having acquired assessment methods and the limit of the volume permitted by nature, it is possible to have sustained usage through administration that extends beyond “borders.” Meanwhile, it is possible to control the “load” with regard to water quality and heat; however, the monitoring of “accumulation” will probably be necessary in the future.

2.6 Conclusion

Under the conditions of increasing population numbers and concentration, as well as globalization, the question becomes one of how to manage the underground environment so as not to damage the future use of the underground environment. It is clear that integrated management is needed to straddle two boundaries: that between the surface and underground, and that between the land and ocean. So long as water continues to circulate in nature, water that has been compartmentalized for societal systems, purposes, and objectives needs to be administered in an integrated and multitiered manner. Another point is the importance of joint management of the underground environment, including switching groundwater from private to public ownership. When undertaken, the perspectives of the “benefits” gained in relationship to those “paid out” will be vital. Last, it is necessary to configure suitable policies that match the capacity/capability of the region and the developmental stage of a city. In humid regions such as Japan in particular, groundwater administration that incorporates not only the concept of it as a “resource” but that of its “circulation” as well, which takes full advantage of the features of groundwater that alleviate change, is necessary.

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Chapter 3
Vadose Zone Hydrology and Groundwater Recharge

Tsutomu Yamanaka

Abstract The rate of recharge (i.e., renewal) of groundwater is a key factor regulating sustainable use of groundwater resources. Hydrological processes in the vadose zone have two important roles: one is to partition the precipitation input into groundwater recharge and evapotranspiration, and the other is to damp down the temporal variability of the recharge flux. Infiltration is the first process after precipitation and is controlled by atmospheric (e.g., rainfall intensity) and soil conditions (e.g., infiltration capacity). Bare soil evaporation and transpiration are the processes of loss of infiltrated water. The rate of bare soil evaporation is sensitive to moisture status in surface soils as well as to atmospheric conditions. Transpiration rate is not very sensitive to these conditions and is capable of taking up water from deeper depths. The temporal and spatial characteristics of the groundwater recharge process change drastically depending on climatic conditions (i.e., humid or arid). Climate change and human activity can affect the rate and temporal variation pattern of groundwater recharge. Proper understanding of vadose zone hydrology is fundamentally important in increasing and improving our knowledge for wise use of groundwater during changes in climate and society.

Keywords Bare soil evaporation • Groundwater recharge • Infiltration • Transpiration • Vadose zone

3.1 Introduction

According to the World Water Assessment Programme (2006), groundwater existing in arid settings of Northern Africa, the Arabian Peninsula, and Australia is considered as a nonrenewable resource. For this type of groundwater resource, the amount of

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storage is fundamentally important for its use. However, in many cases over the world, groundwater resources are renewable. Indeed, even the nonrenewable groundwater was derived from recharge events under paleoclimatic conditions and is still recharged very slowly under the current climate condition. Thus, the rate of renewal or recharge is a key factor that regulates sustainability of groundwater use.

Recharge of groundwater usually occurs from precipitation through the vadose zone. The *vadose zone* is defined as the depth zone between the ground surface and the water table. In the vadose zone, the pressure head is less than atmospheric pressure. Most parts of the zone are unsaturated, whereas its lowest part (i.e., the capillary fringe) is saturated.

The vadose zone has two important roles in recharging groundwater: one is to partition the precipitation input into recharge flux as output to groundwater body and evapotranspiration flux as loss, and the other is to damp down the temporal variability of the output. In particular, the former role is crucial in regulating the amount of groundwater recharge and thus the amount of groundwater that can be used sustainably by human beings, and the latter role is capable of affecting the former role through interaction between them. Therefore, understanding of vadose zone hydrology is effective for better management of groundwater resources.

In this chapter, the phenomenological aspects of individual processes of vadose zone hydrology, such as infiltration, bare soil evaporation, and transpiration (including root water uptake), are briefly explained in terms of groundwater recharge. Moreover, unique characteristics of groundwater recharge under both humid and arid climates are described as well as their relationship to climate change and human activities. For further reading, the work of Stephens (1996) and Parlange and Hopmans (1999) is recommended. Theoretical aspects of vadose zone hydrology are also described in the literature of soil physics (Hillel 1980; Campbell 1985) or hydrology (Ward and Robinson 1990; Brutsaert 2005).

3.2 Infiltration

The term *infiltration* is used to describe the process of water entry into the soil across the ground surface (Ward and Robinson 1990). The maximum rate of infiltration is called the infiltration capacity, and it depends on soil hydraulic properties. It is observed that the infiltration capacity varies with time during a storm event, as schematically illustrated in Fig. 3.1.

In small or moderate storm events where rainfall intensity is less than the infiltration capacity, actual infiltration rate equals to the rainfall intensity. This situation may be regarded as the atmospheric control phase of the infiltration process. On the other hand, in greater storm events where rainfall intensity exceeds the infiltration capacity, actual infiltration does not depend on rainfall intensity but equals the infiltration capacity. This stage corresponds to the soil control phase.

In the soil control phase, overland flow is generated by excess water (i.e., precipitation minus infiltration), even if the water table does not rise to the ground surface. However, this type of overland flow (i.e., so-called Hortonian

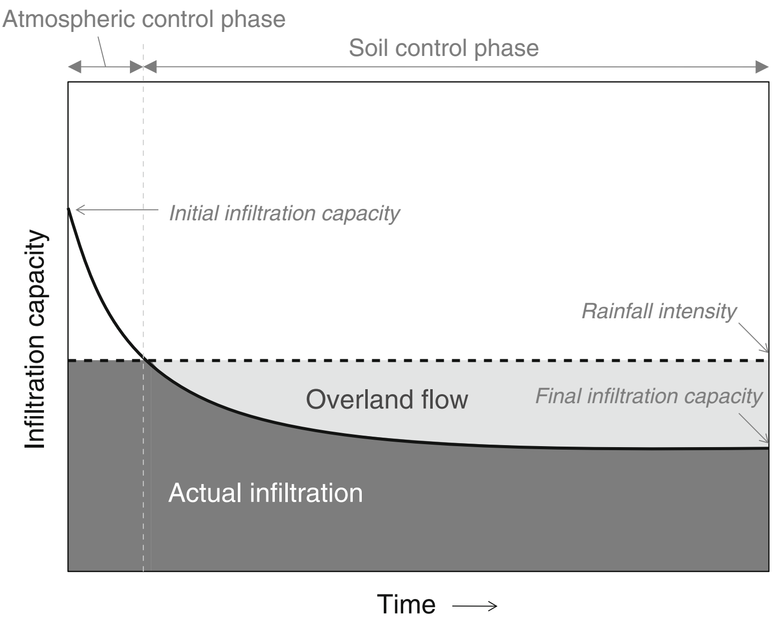

Fig. 3.1 Schematic illustration of temporal variation of infiltration capacity

overland flow) is uncommon for well-vegetated temperate areas (except for some poorly managed forest floors; Tsujimura et al. 2006), because the infiltration capacity in such areas is generally greater than possible rainfall intensity (e.g., 100 mm/h). Consequently, most of the precipitated water enters below the ground surface, and cumulative infiltration flux is basically regulated by the amount of precipitation.

When a large amount of precipitated water enters into a dry soil, a saturated or nearly saturated zone can be formed just below the ground surface. If the water input continues, the zone will expand downward. Its bottom boundary, where the moisture gradient is very steep, is called the wetting front. When water enters into a relatively moist soil or the precipitation rate is relatively small, the wetting front tends to be unclear.

Downward movement of the wetting front does not occur uniformly in space. Micro-relief of the ground surface or spatial heterogeneity in soil hydraulic properties induces a concentrated, preferential flow and makes the wetting front spatially unstable (e.g., fingering; Fig. 3.2a). Within the saturated zone (exactly speaking, a positive pressure zone), the preferential flow can occur through macropores (Fig. 3.2b). In some cases, the preferential flow contributes significantly to downward transport of water and thus to groundwater recharge.

In forests, stemflow sometimes induces concentrated flow around the tree. Although the amount of stemflow (S_f) is usually no more than 10 % of gross rainfall (R_g), the infiltration rate within a limited area affected by the stemflow can be remarkably higher than that of throughfall. As a result, stemflow-induced

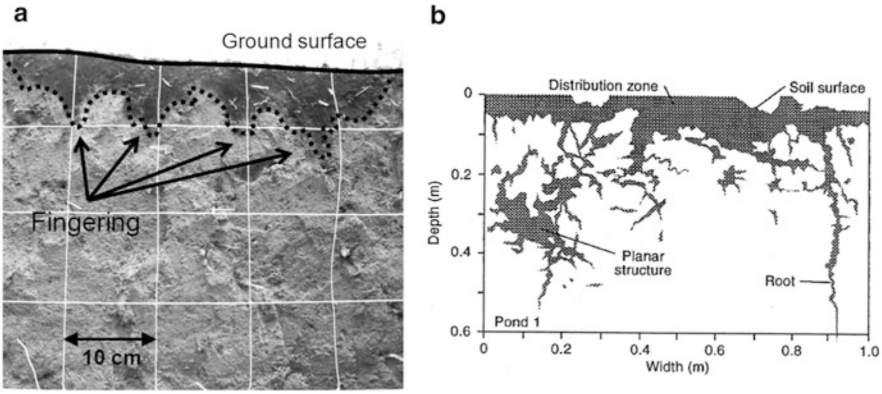

Fig. 3.2 Results of dye tracer experiment demonstrating (**a**) fingering (modified from Iida and Hamada 2009) and (**b**) macropore flow (after Scanlon et al. 1997)

Fig. 3.3 Photograph of inhomogeneous moisture distribution caused by stem flow-induced infiltration. (Modified from Tanaka et al. 2004)

infiltration can penetrate into the deeper depths and thus contribute to effective recharge of groundwater (Fig. 3.3). According to a case study of a pine forest by Taniguchi et al. (1996), the recharge component originated from stemflow reached 10–20 % of total recharge despite the low S_f/R_g (approximately 1 %).

3.3 Bare Soil Evaporation

A part of the infiltrated water returns to the atmosphere via either bare soil evaporation or transpiration and never contributes to groundwater recharge. In other words, only residual water that escapes opportunities of evapotranspiration can contribute to groundwater recharge. Although it is not easy to measure evaporation and transpiration separately, the two are clearly different processes.

Bare soil exists in part even in vegetated lands as gaps between plants. For vegetated lands, the fraction of evaporation from the soil surface against total evapotranspiration is estimated to be about 10 % or less. On the other hand, large-scale bare soil can be found as deserts, sand dunes, and playas.

It is well known that process of bare soil evaporation has an atmospheric control phase and a soil control phase. Evaporation rate equals atmospheric demand when the soil is wet enough and becomes suppressed during soil drying (Fig. 3.4). The former corresponds to the atmospheric control phase, and the latter is the soil control phase.

The atmospheric demand for evaporation or maximum evaporation rate under given atmospheric conditions is called the potential evaporation. The so-called Penman equation represents a relationship between the potential evaporation rate and atmospheric conditions, suggesting that potential evaporation is mainly controlled by four factors: net radiation, wind speed, vapor pressure deficit, and air temperature.

As the soil dries, the dry surface layer (DSL) forms, in which water content is very small and liquid water movement hardly occurs (Fig. 3.5). Under such a situation, evaporation of water occurs at the bottom boundary of the DSL rather than at the soil surface, and water moves upward through the DSL in vapor phase principally by molecular diffusion. Molecular diffusion of water vapor in the

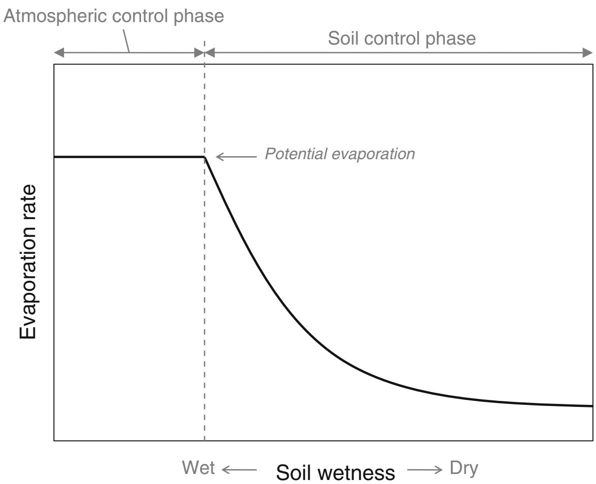

Fig. 3.4 Schematic illustration of the relationship between evaporation rate and soil wetness

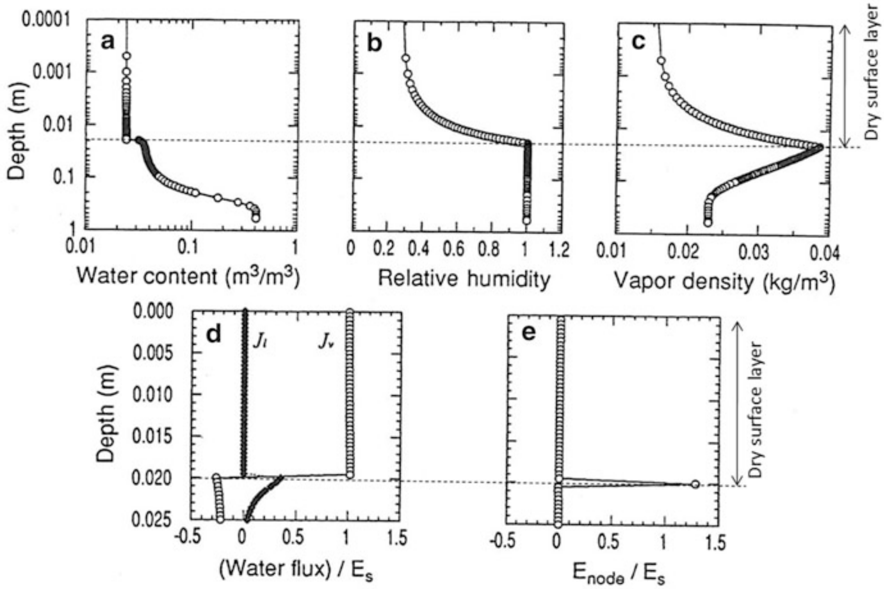

Fig. 3.5 Numerical simulation results of vertical distribution of (**a**) water content, (**b**) relative humidity, (**c**) vapor density, (**d**) water fluxes normalized by evaporation flux (E_s) in liquid phase (J_l) and vapor phase (J_v), and (**e**) evaporation density at a calculation node (E_{node}) normalized by E_s. (Modified from Yamanaka et al. 1998)

soil occurs very slowly as compared to the turbulent transport in the atmosphere. In other words, development of the DSL increases resistance to water vapor transport from the soil to the atmosphere; this is the reason why the rate of bare soil evaporation is suppressed during soil drying.

The DSL is formed more easily in coarse-textured (i.e., sandy) soils than in fine-textured soils, because the change in unsaturated hydraulic conductivity caused by soil drying is more drastic (Fig. 3.6). Therefore, bare soil evaporation from sandy soils decreases quickly after storm events; that is, sandy soils tend to save soil moisture and contribute to groundwater recharge. Gravel mulch acts to enhance formation of localized DSL and an dramatic increase in resistance to water vapor transport to the atmosphere (Yamanaka et al. 2004). For this reason, gravel mulch can reduce evaporation rate and increase groundwater recharge.

If the water table is situated at a shallow depth in fine-textured soils, water required for evaporation can be supplied continuously from the water table through the capillary fringe. Such a situation often occurs at the discharge zone (e.g., playas and riparian zones) in the regional groundwater flow system. In this case, the cumulative amount of evaporation can exceed that of precipitation under arid climates and no net recharge of groundwater may occur.

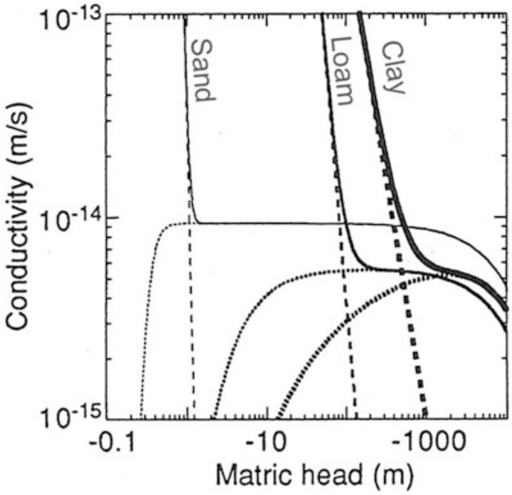

Fig. 3.6 Dependence of liquid (*dashed line*), vapor (*dotted line*), and total (*solid line*) conductivities on matric head for sand, loam, and clay. (Modified from Yamanaka et al. 1998)

3.4 Transpiration

Transpiration from vegetation also has atmospheric and soil control phases. When soil moisture is abundant, the transpiration rate nearly equals atmospheric demand. However, it is reduced during soil drying. The switching between the two phases is less clear than in bare soil evaporation.

The reduction of transpiration directly results from stomatal closure, which increases resistance to water vapor transport from the stomatal cavity to the atmosphere. The stomatal closure is caused by a decrease in leaf water potential, and the leaf water potential is determined by a balance of transpiration rate and root water uptake rate in the soil-plant-atmosphere continuum (SPAC).

The SPAC is analogous with an electric circuit (Fig. 3.7). Water flows from the soil with higher potential to the atmosphere with lower potential. In the root system as a parallel circuit, water flows through a set of root paths so as to minimize resultant resistance in the system. In many cases, root density is higher at shallower depths. Because microscopic resistance to water transport is less in plant roots than in soil matrix, the bulk resistance for a depth zone is smaller at shallower depths. For this reason, even if the hydraulic potential (i.e., pressure potential plus gravity potential) is uniform over the soil profile, plants take up water mainly from shallower depths (Fig. 3.8a).

As soil water potential decreases in shallower depths, the hydraulic gradient (i.e., difference in hydraulic potential) between the shallow zone and the atmosphere becomes smaller. On the other hand, water potential in deeper depths remains higher, so that water uptake rate increases relatively in deeper depths and decreases

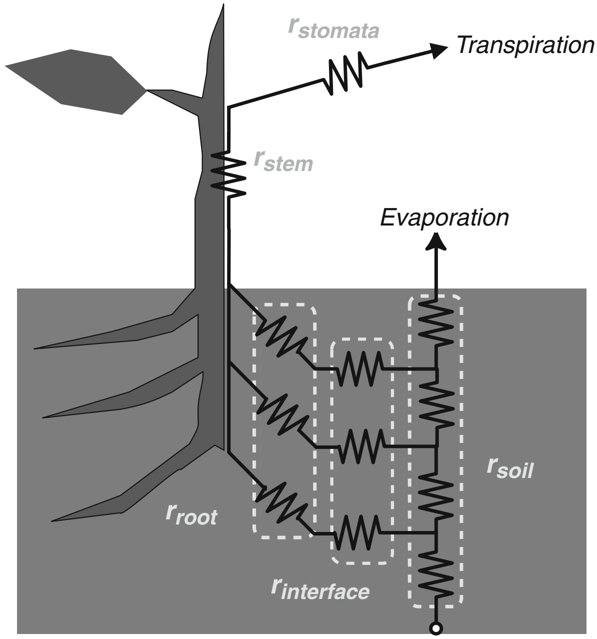

Fig. 3.7 Schematic illustration of water transport in the soil-plant-atmosphere continuum (SPAC) analogous with an electric circuit, which is composed of resistance at the soil (r_{soil}), soil–root interface ($r_{\mathrm{interface}}$), root (r_{root}), stem (r_{stem}), and stomata (r_{stomata})

in shallower depths if transpiration rate is high (Fig. 3.8b). In this way, plants can take up water from relatively moist layers in the soil profile, and transpiration rate meets atmospheric demand even though the surface soil dries. However, because soil water steadily decreases because of root water uptake, mean soil water potential over the soil profile and leaf water potential also decrease. When leaf water potential is less than a threshold value, transpiration becomes suppressed, which is the mechanism of switching from the atmospheric control phase to the soil control phase.

Plants of some species (e.g., sugar maple, gambel oak, pigeon pea) take up water from moist deep soils and release it to dry shallow soils. This phenomenon is called *hydraulic lift*. Hydraulic lift occurs in such plants during the nighttime to store water temporarily in shallow soils, and then in the daytime the plants utilize the lifted water for transpiration (Fig. 3.9). In some cases, understory plants also depend upon the lifted water for surviving during severe droughts; the hydraulic lift acts as a natural pump.

As explained in Sect. 3.2, plants contribute in part to groundwater recharge via stemflow-induced infiltration. However, in most cases, plants gain opportunities for water loss from the vadose zone and thus reduce rather than enhance groundwater recharge.

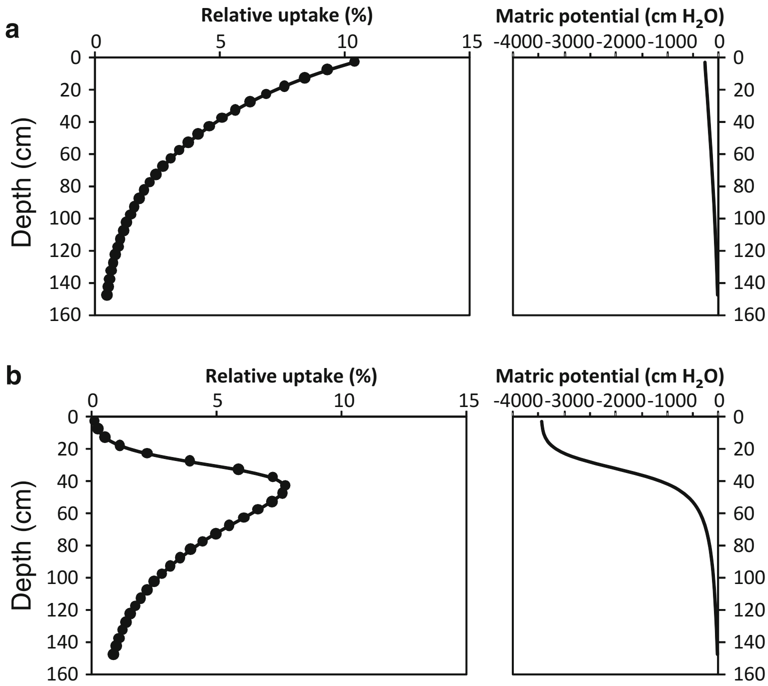

Fig. 3.8 Numerical simulation results of depth profile of (*left*) root water uptake from 5-cm-thick layers relative to total uptake (i.e., transpiration) and (*right*) matric water potential for a wet case (**a**) and a dry case (**b**)

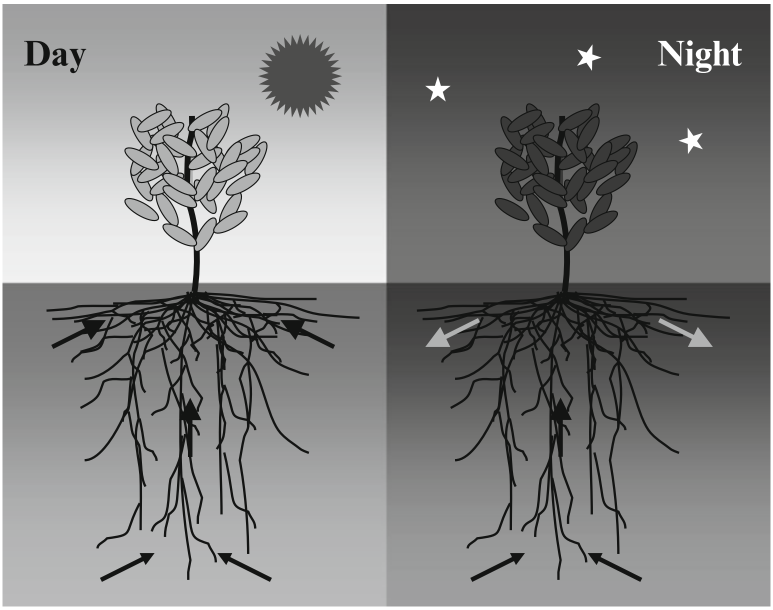

Fig. 3.9 Schematic illustration of water flow through root system associated with hydraulic lift

3.5 Groundwater Recharge Under Humid Climate

Under a humid climate, annual precipitation is generally greater than annual evapotranspiration, so that a certain amount of groundwater recharge is usually expected. Table 3.1 summarizes groundwater recharge rates observed in previous studies in various locations, suggesting that recharge rate and the recharge-to-precipitation ratio depend mainly on annual precipitation.

The total amount and temporal variation pattern of groundwater recharge are affected by soil type as well as amount of precipitation. In sandy soils, which have low water retention, downward soil water movement and groundwater recharge soon occur in response to each storm event (Fig. 3.10). On the other hand, in fine-textured soils, the response of groundwater recharge to each storm event tends to be unclear and the temporal variation of groundwater recharge flux becomes moderate, because fine-textured soils with high water retention have a large capacity for water storage.

Under humid climates, there can be several different pathways of recharge other than simple infiltration of precipitation. For example, rice paddy fields are often situated on the vadose zone, and the consequent water leakage can contribute to groundwater recharge through the vadose zone. Effluent seepage from rivers, as well as leakage from paddy fields, is also an important contributor to groundwater recharge, particularly at recharge zones in the regional groundwater flow system. Using an isotopic tracer technique, Wakui and Yamanaka (2006) have evaluated the contribution from different pathways in an alluvial fan as a case study (Fig. 3.11). According to their results, which may be common to other alluvial fans, effluent seepage from rivers is the principal contributor in areas just adjacent (~2 km) to the rivers, whereas in other areas simple infiltration of precipitation is important. Groundwater recharge from paddy fields was minor in the region, although it should be noted that the relative importance of each contributor depends on conditions of

Table 3.1 Observed values of groundwater recharge rate and related information

References	Location	Soil type	Recharge (mm/year)	Precipitation (mm/year)	Recharge/precipitation
Shimada (1988)	Kanagawa, Japan	Loam	913	1,672	0.55
Kayane et al. (1980)	Tokyo, Japan	Loam	885	1,550	0.57
Yamanaka et al. (2005)	Ibaraki, Japan	Loam	399	1,321	0.30
Yamanaka et al. (2005)	Hiroshima, Japan	Sand	523	1,262	0.41
Anderson and Sevel (1974)	Denmark	Till	358	780	0.46
Allison and Hughes (1974)	Australia	Sandy loam	40–140	750	0.05–0.19
Sukhija and Shah (1976)	India	Sand, loam	15–56	700	0.02–0.08
Vogel et al. (1974)	South Africa	Sand	10	500	0.02

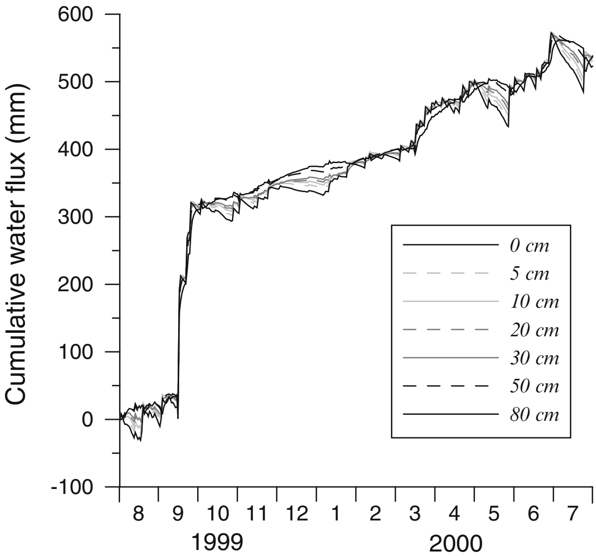

Fig. 3.10 Temporal evolution of cumulative water flux at different depths for a sandy soil site. (Modified from Yamanaka et al. 2005)

topography, hydrogeology, and land use patterns. This kind of information allows us to discuss influences of water use or land use on groundwater resources. For example, if upstream river water is regulated by dams or taken for agricultural use or drinking water supply, groundwater recharge from rivers would decline at areas near the rivers. On the other hand, at inter-river areas, if landfill sites are constructed or much agricultural chemical/fertilizer spraying occurs, groundwater will be placed at high risk of pollution.

Future climate (e.g., precipitation and air temperature) change is capable of changing groundwater recharge rate principally through atmospheric control phases of infiltration and evaporation. The Intergovernmental Panel on Climate Change (IPCC) reported that in many regions with a humid climate both precipitation and evapotranspiration will increase and the increase in precipitation is likely greater (Solomon et al. 2007; Parry et al. 2007). Therefore, groundwater recharge under a humid climate will increase to some extent. Although there have been few studies on the effect of climate change on groundwater recharge, including groundwater–surface water interaction (Alley 2001), changes in river level caused by climatic change can influence not only groundwater levels but also groundwater recharge (Allen et al. 2003). In addition, climate change may induce vegetation changes that affect groundwater recharge, for instance, through evapotranspiration or stemflow-induced infiltration.

Land use or land cover change caused by human activity can also modify groundwater recharge. Unfortunately, it is impossible to predict any changes in groundwater recharge from land use and land cover changes because there is no reliable scenario for such changes. However, the isotopic tracer technique mentioned here would be useful to assess the human impact on groundwater recharge in the area in question.

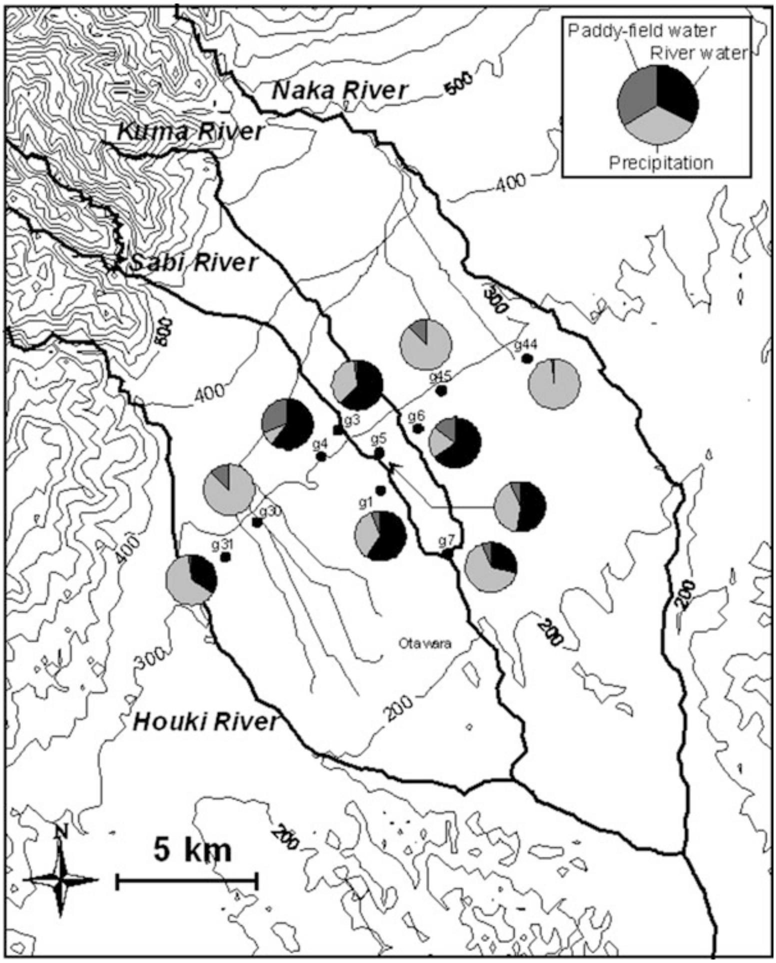

Fig. 3.11 Spatial distribution of the contribution ratio from different groundwater-recharge pathways, estimated for a alluvial fan using an isotopic tracer approach. (Modified from Wakui and Yamanaka 2006)

3.6 Groundwater Recharge Under Arid Climate

From the aspect of spatial structure, groundwater recharge has two modes: one is the diffuse recharge occurring homogeneously in space, and the other is the concentrated recharge that occurs only at limited areas (Gee and Hillel 1988). Under arid climates, this distinction is crucial.

A review by Scanlon et al. (1997) of downward water flux at depths deeper than the root zone at various arid settings throughout the world suggests that a negligibly small amount of diffuse recharge of groundwater is commonly found at vegetation-covered

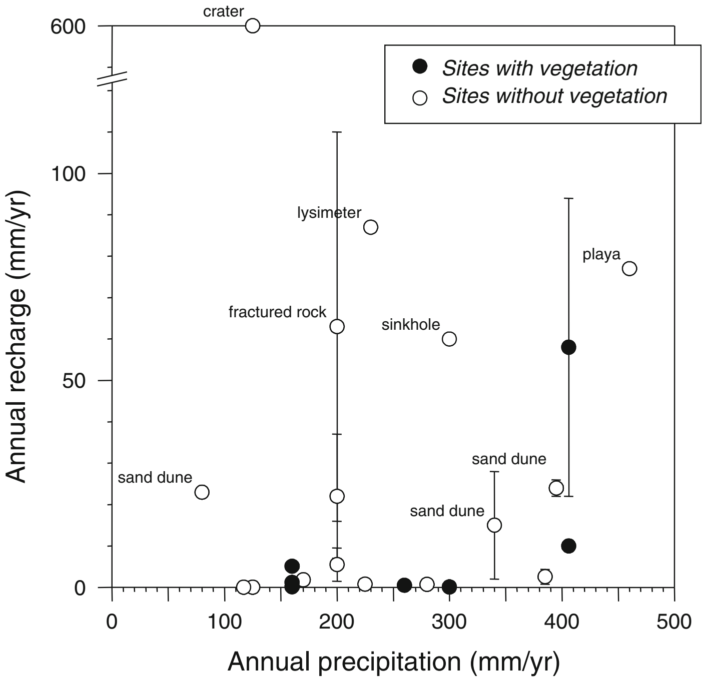

Fig. 3.12 Relationship between annual precipitation and annual recharge at various arid settings. (Data from Scanlon et al. 1997)

lands with precipitation no greater than 300 mm/year (Fig. 3.12). Under such conditions, annual precipitation nearly balances annual evapotranspiration (Fig. 3.13). However, even if precipitation is less than 300 mm/year, a considerable amount of concentrated recharge can occur at limited areas (Scanlon et al. 2006), such as depressions (e.g., sinkholes, playas, craters), outcrops of fractured rock, and highly permeable sand dunes (Fig. 3.12).

Because diffuse recharge tends to decrease with increasing aridity, the concentrated recharge occurring in limited areas is of major importance in arid zones (Gee and Hillel 1988). Preferential flow, such as macropore flow in fractured rocks or fissured sediments, is particularly important with respect to groundwater recharge and contaminant transport in arid settings (Scanlon et al. 1997). To assure sustainable use of renewable groundwater, areal evaluation of the amount of concentrated recharge over a drainage basin is required, although at present it is difficult to identify the occurrence of the concentrated recharge and to spatially aggregate it.

Arid and semiarid climates are characterized by rainfall with high temporal and spatial variability and low frequency of occurrence. Interannual or interdecadal variability of rainfall remarkably affects groundwater recharge in arid and semiarid regions. Kearns and Hendrickx (1998) performed a 100-year simulation of groundwater recharge in southern New Mexico, where mean annual rainfall is 203 mm,

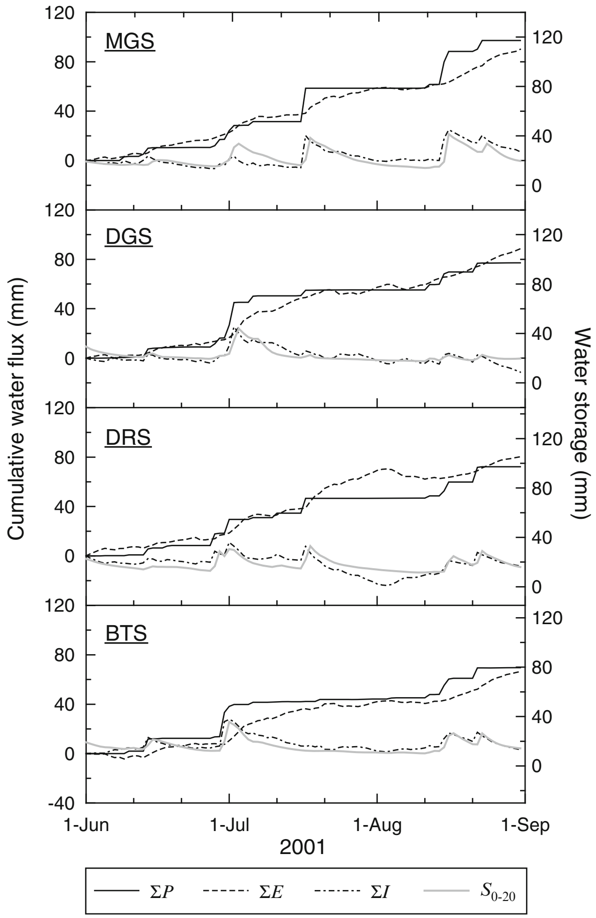

Fig. 3.13 Temporal evolution of cumulative precipitation (ΣP), evapotranspiration (ΣE), net infiltration ($\Sigma I = \Sigma P - \Sigma E$), and soil water storage in the 0–20 cm depth zone (S_{0-20}), observed at four sites (*MGS* Mandalgobi, *DGS* Delgertsogt, *DRS* Deren, *BTS* Bayan tsagaant) in Mongolia. (After Yamanaka et al. 2007)

and showed that for a grass-covered loamy fine sand, the recharge rate is normally approximately 2 mm/year whereas it exceeds 25 mm/year in a year in which the largest daily rainfall of 165 mm was recorded.

As mentioned in the previous section (3.5), global warming can induce increases of both precipitation and evapotranspiration. As for arid regions, groundwater recharge is likely to increase in some areas such as Sahel, the Near East, northern China, and the western USA, and decrease in some other areas such as northeastern Brazil, southwest Africa, and along the southern rim of the Mediterranean Sea (Parry et al. 2007). IPCC reported that global warming will accompany or is accompanying extremely heavy rainfall in arid regions as well as extremely intense drought (Solomon et al. 2007). Such a heavy rainfall can effectively contribute to diffuse recharge even if its frequency is low. In addition, concentrated recharge will be more effectively enhanced by the heavy rainfall. On the other hand, climate change potentially affects vegetation condition and then groundwater recharge. However, it is not uncertain whether increase in precipitation enhances vegetation growth with more evapotranspiration resulting, thus introducing no change in groundwater recharge. The effect of land use/cover change by human activity may not be as significant in arid regions compared to humid regions, although agricultural water use can contribute to groundwater recharge to some extent.

3.7 Concluding Remarks

Groundwater recharge rate is one of the most important factors regulating the sustainability of groundwater use. Climate change and human activity affect the rate and temporal variation pattern of groundwater recharge. Our current knowledge for future projection is, unfortunately, not sufficient. Proper understanding of vadose zone hydrological processes that are directly or indirectly linked to groundwater recharge is fundamentally important in increasing and improving our knowledge for the wise use of groundwater during changes in climate and in society.

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Chapter 4
Use of Water Quality Analysis for Groundwater Traceability

Takanori Nakano

Abstract Water contains various elements with different concentrations. The quality of terrestrial water varies spatially and temporally in accordance with the amount and quality of precipitation, geology, and artificial materials in the watershed. The overall characteristics of water quality are visualized as a hexadiagram, which is expressed by the concentration and proportion of the major cations and anions. Water quality of groundwater displays geographic variation in a basin, which is a function of the dissolution and precipitation of minerals, sorption–desorption at mineral surfaces, ion exchange with clays, and oxidation–reduction reactions in the aquifer. Stable isotopes of hydrogen and oxygen in groundwater are useful to identify the recharging area, whereas those of sulfate and nitrate are useful to evaluate the degree of oxidation–reduction resulting from bacterial activity. The strontium isotope ratio is known to be a powerful hydrogeological tracer, because the geochemical behavior of Sr is similar to that of Ca, which is the dominant cation derived from rocks. The geochemical and isotopic composition of water and its various constituents can be used as traceability data that reproduce the natural environment and human activities encountered along the path of groundwater.

Keywords Chemical weathering • Hexadiagram • Reduction • Stable isotopes • Water quality

4.1 Introduction

If a water-related problem such as pollution or water shortage arises in a river or stream, the problem source can be found by working one's way upstream. However, as groundwater flows through sediment and rock, it is nearly impossible to identify

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problem sources in the same manner. Moreover, human activity impacts not only surface water but also groundwater in a variety of ways, and with increasing urbanization and population growth, the demand for consistent, high-quality groundwater continues to increase. Traceability refers to the capability to trace a material back to its source. The implementation of traceability labels (historical information) on food and other products, which enables the tracking of a product's history from manufacturing, through processing, packaging, distribution, and, ultimately, the retail market, is an example of the rapid adoption of an information system to ensure the safety of and consumer confidence in man-made products. To ensure the safe and sustainable use of groundwater, the development of analogous methods for achieving groundwater traceability—from its point of origin, through underground flow, and, ultimately, to its reemergence in rivers or other bodies of water—is becoming increasingly critical.

Water contains a variety of constituents, the concentration and relative proportions of which are collectively called "water quality." Water quality analysis typically entails the measurement of a specific set of parameters, such as nitrate, heavy metals, and organic compounds. However, water and its various constituents can be used as traceability data that contain a record of the natural environment and human activities encountered along its path (Fig. 4.1). For example, groundwater retains qualities of the water with which it is recharged (for example, from rivers), but is also altered by reacting with the rocks and sediments through which it has flowed and by mixing with water from different sources. Such "reactions" with sediments, of course, comprise multiple processes, including the dissolution of minerals, adsorption–desorption at mineral surfaces, ion exchange with clays, and oxidation–reduction reactions. The groundwater is further altered when by-products of human activity (agriculture, manufacturing, etc.) are transported into the ground by percolating water.

However, every reaction and human activity has a characteristic impact on specific components of water quality. Thus, by analyzing multiple components of water quality as well as the ensemble of components, it should be possible to identify the source of the water as well as the various processes encountered along its path. In the majority of cases, groundwater is connected to rivers or streams, and by comparing the quality of groundwater and surface waters, it is possible to identify the relationship between the two. Recent years have seen the development of methods to analyze numerous elements using stable isotopes. Stable isotopes are an element's chemical fingerprint and can provide reliable information regarding the origin of a water sample and the elements dissolved therein.

For such water quality information to become widely used as tools for water management, the government and the public, along with other stakeholders in the water environment, need to understand the utility of such data in achieving "traceability." To that end, it is also necessary to present water quality data and to describe the overall characteristics of water quality in an easy-to-understand format. In this chapter, I introduce research and methods developed for visualizing groundwater quality based on the analysis of two types of geochemical data, the elemental content and stable isotope ratio of water, from the standpoint of achieving traceability.

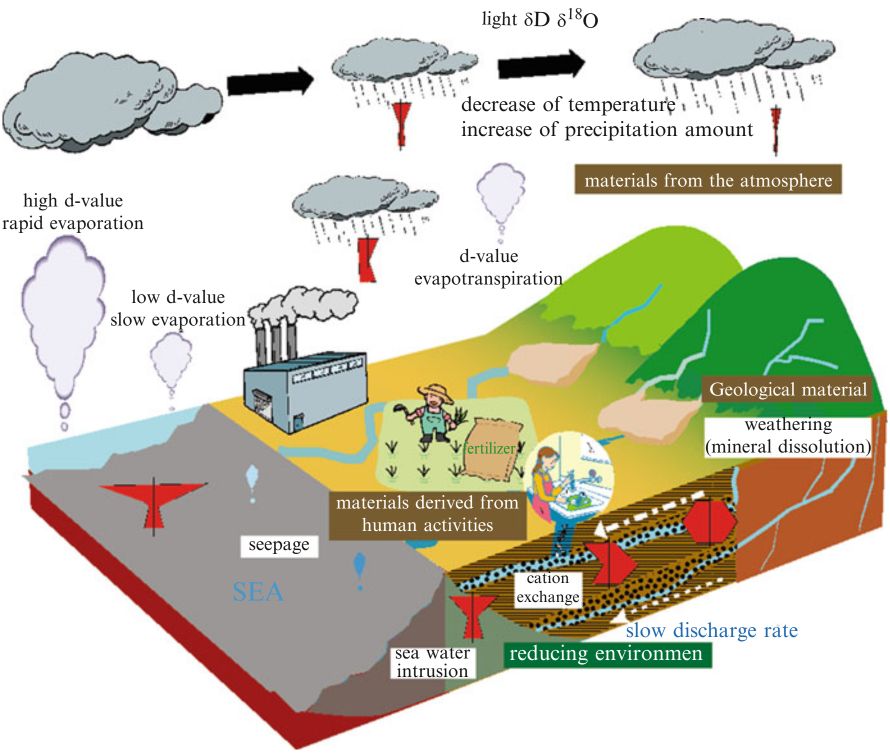


Fig. 4.1 Impacts of the natural environment and human activity on water quality within the hydrologic cycle

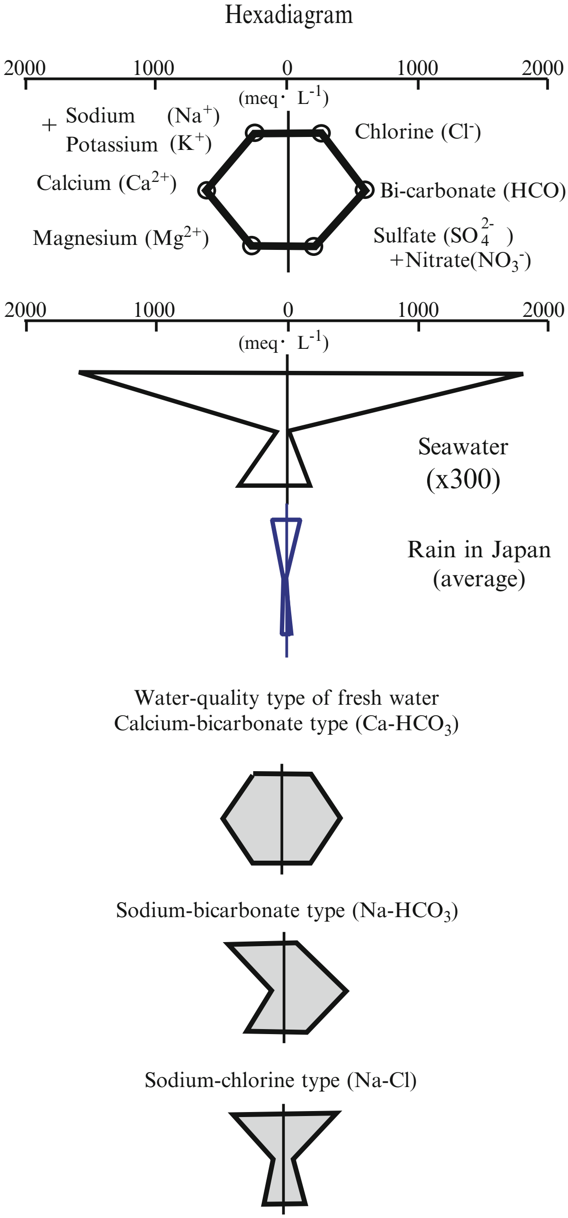
4.2 Water Quality Index

4.2.1 Hexadiagrams

Water quality varies greatly depending on the form of water (rain, river, ocean) as well as location and time. A number of methods have been proposed for presenting water quality in a readily understandable format. Chemical elements dissolved in water exist either as positively charged cations or negatively charged anions. The major cations found in water are sodium (Na^+), potassium (K^+), calcium (Ca^{2+}), and magnesium (Mg^{2+}); the major anions include chloride (Cl^-), bicarbonate (HCO_3^-), sulfate (SO_4^{2-}), and nitrate (NO_3^-). Accordingly, water quality can be characterized by the concentration and relative proportions of these eight cations and anions.

Among the major cations, potassium and sodium are typically grouped together because potassium and sodium are relatively strongly correlated, both being alkaline elements, and also because potassium is found in relatively low concentrations in water. Among the major anions, nitrate and sulfate are generally grouped

Fig. 4.2 Hexadiagrams and types of freshwater



together, because both typically originate from fertilizer, household wastewater (grey water), and other human activity, and nitrate is found in low concentrations. The cations and anions are thus each divided into three components. As illustrated in Fig. 4.2, plotting the cationic components to the left and the anionic components to the right and joining the plotted points results in a six-sided figure, with the distance of each point from the origin indicating concentration. Such a plot is known as a hexadiagram. The overall concentration of anions and cations are the same because the overall charges must balance. A hexadiagram can thus be used to capture the overall characteristics of water quality.

4.2.2 Formation of Water Quality Through Chemical Weathering of Minerals

Water in which the concentrations of sodium and chlorine dominate that of other ionic species is designated as “sodium-chlorine type” water (Fig. 4.2). Seawater is a prime example of this water type. In contrast to seawater, the quality of freshwater on land varies greatly depending on geographic location. One reason for this lies in the variability of natural factors in different watersheds. In particular, precipitation and local geology greatly impact freshwater quality. Rain and snow (generally referred to as “wet deposition”), which result from the condensation of water vapor around aerosols, that is, minute particles suspended in the air, typically contain little in the way of dissolved minerals. That is, they are very close to pure water and, as such, their hexadiagrams are small. However, on closer examination, it becomes apparent that precipitation quality varies by location. For example, the atmosphere in the vicinity of ocean islands and coastal areas is rich with aerosols generated by the evaporation of ocean spray, and, as such, the precipitation consists of sodium-chlorine type water. In contrast, in urban and inland areas, the atmosphere is richer in calcium-containing aerosols, and, thus, the precipitation has a higher calcium content.

In regions that experience heavy precipitation, precipitation significantly impacts the water quality of both surface water and groundwater. Generally speaking, water in areas of heavy precipitation such as mountains exhibits the lowest dissolved mineral content. Precipitation is naturally acidic and reacts with the soil and rock, dissolving the minerals contained therein as it percolates into the ground. Such reactions are referred to as chemical weathering. Although weathering results from the action of all kinds of acids, the most common naturally occurring acids include carbonic acid (H_2CO_3), sulfuric acid (H_2SO_4), nitric acid (HNO_3), hydrochloric acid (HCl), and hydrofluoric acid (HF). In soils, decomposition of organic matter also generates organic acids. Among the aforementioned acids, carbonic acid is the most important when considering precipitation quality. Carbonic acid is generated when atmospheric carbon dioxide is dissolved in water according to the following reaction:

$$H_2O + CO_2 \Rightarrow H_2CO_3 \Rightarrow H^+ + HCO_3^- \tag{4.1}$$

As can be seen in chemical equation (4.1), acidity is generated by the release of H^+ ions. Rocks are made up of combinations of minerals having varying chemical compositions. It is the H^+ ions, common to all acids, that promote the dissolution of various minerals. Chemical weathering is a neutralization reaction in which part of an acid is converted into an anion and the constituents of the dissolved mineral remain in solution, in most cases, as cations. As such, the more a water has experienced chemical weathering, the higher its overall cation and anion content.

A mineral's reactivity with water depends on its specific chemical composition. In general, minerals with high calcium content are readily weathered by carbonic acid, resulting in water with high concentrations of both Ca^{2+} and HCO_3^-. For example, the chemical weathering by carbonic acid of limestone used in cement proceeds in the following manner:

$$CaCO_3 + H_2CO_3 \Rightarrow Ca^{2+} + 2HCO_3^- \quad (4.2)$$

Much of the earth's rock is composed of a variety of silicate minerals, which differ in their relative calcium and silicon content, as well as the content of other elements. Chemical weathering of the various calcium-containing silicates is often represented by the following weathering reaction of the group's simplest example, $CaSiO_3$:

$$CaSiO_3 + H_2O + 2H_2CO_3 \Rightarrow H_4SiO_4 + Ca^{2+} + 2HCO_3^- \quad (4.3)$$

Freshwater on land is greatly influenced by chemical weathering and, in general, contains high concentrations of the cation Ca^{2+} and the anion HCO_3^-; such water is designated as a "calcium-bicarbonate ($Ca^{2+} - HCO_3^-$) type" water (Fig. 4.2). In reality, minerals containing Na, K, and Mg also undergo chemical weathering, and the concentrations of these cations also vary. For example, although the granite prevalent in continental rock contains high levels of Na and K, rocks in volcanic regions contain high levels of Mg; these differences are also reflected in the water quality of the respective regions. A watershed's geologic environment greatly influences the quality, and especially the cation profile, of water in the region.

4.2.3 Geologic Characteristics Impacting the Quality of Floodplain Groundwater

Although precipitation quality can vary significantly from event to event and from season to season, these variations disappear upon percolation, and the quality of groundwater or subsurface runoff (base flow) is stable. Groundwater quality, particularly that of deep strata, changes very little over time. In contrast, as already discussed, terrestrial water is impacted by the amount and quality of wet deposition, the rock and soil through which it passes, as well as by a variety of human activities; as such, it varies significantly by geographic location. By comparing the characteristics of groundwater in different locations, it is possible to delineate aquifer recharge areas, and by identifying the factors that determine geographic variability, to understand the impact of these factors on groundwater quality.

However, the same groundwater, which in mountainous areas flows along the soil–bedrock boundary (subsurface flow) or through rock fractures (fissure water), in floodplains flows through highly permeable strata consisting of sand or gravel (Fig. 4.1). As expected, these differences in flow environment are strongly reflected in water quality. The strata that make up floodplains comprise deposits of sediment generated by the weathering of upstream rock and consist primarily of silicate minerals. The clay sediments resulting from weathering are relatively insoluble in acid and contain little calcium. In other words, floodplain sediments are relatively less soluble than the rocks found in mountainous areas. The clay minerals and organic matter in floodplain sediments have high ion-exchange and adsorptive capacity, and the slow-flowing groundwater environment can easily become reducing.

In addition to these factors, the strata of coastal plains are often significantly influenced by seawater. After strata are formed, their mineral and chemical composition is altered by reaction with interstitial water. This process, called diagenesis, results in authigenic (diagenetic) minerals. Marine sediments, in particular, contain authigenic sulfide minerals that, when oxidized, become strongly acidic, and the interstitial water retains constituents of seawater from their time of deposition. The presence of these authigenic minerals and adsorbed seawater constituents greatly alters the groundwater quality.

4.2.4 Changes in Cation Content Associated with Groundwater Movement: The Aquifer in the Northeast of the Osaka Plain

The area between Shimamoto and Takatsuki in the northeast region of the Osaka Plain is known for its abundant groundwater supply. Figure 4.3 shows hexadiagrams of the water quality of the artesian aquifer at depths to 100 m (Yamanaka et al. 2005). The rivers (such as the Akuta and Hio Rivers) originating in the mountainous region to the north flow southward across the plain and merge with the southwestwardly flowing Yodo River before empting into Osaka Bay. In the mountainous region, the river water is of the calcium-bicarbonate type, whereas the water of the Yodo River is of the sodium-bicarbonate type.

Meanwhile, it is evident that there is geographic variability in groundwater quality. Although bicarbonate is the dominant anion in all regions, the groundwater contains high levels of calcium in the north, high levels of sodium in the south, and relatively high levels of magnesium in between. Thus, it is possible to divide the region into three zones based on groundwater quality, starting with the calcium-bicarbonate type in the north, the magnesium-bicarbonate type in the middle, and the sodium-bicarbonate type in the southwest. Based on the fact that approximately 30,000 t groundwater is pumped out from the southwest portion of this region on a daily basis,

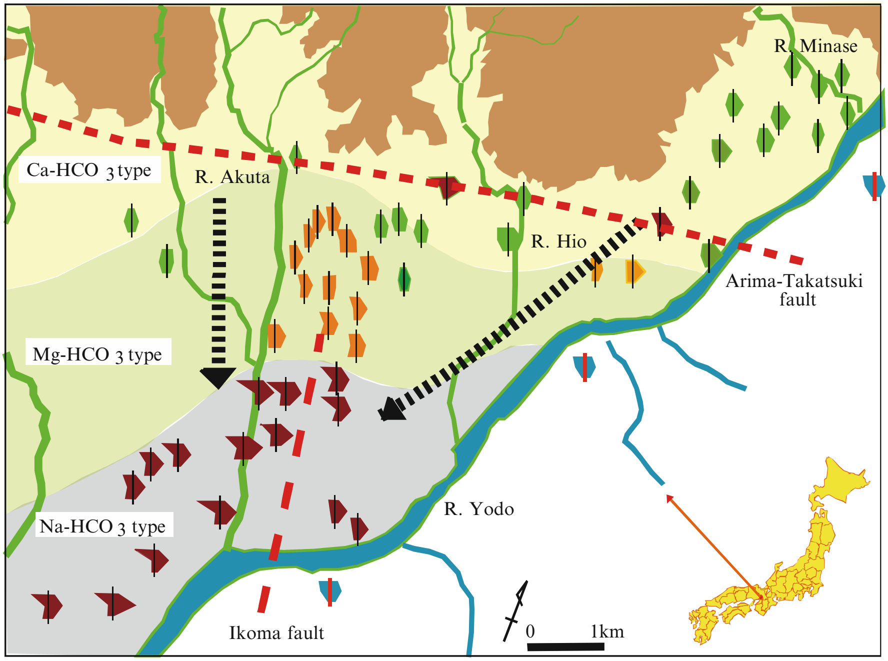


Fig. 4.3 Hexadiagrams of groundwater in northeastern Osaka City. *Dashed line* indicates the direction of groundwater flow. (Modified from Yamanaka et al. 2005)

it is believed that the groundwater flows in a southwesterly direction. Accordingly, it is believed that the geographic variability in groundwater quality is a function of the groundwater movement. What is of interest is that, despite the change in water quality, the total concentration of dissolved ions changes little with location. These observations indicate that the groundwater experiences cation exchange as it passes through the aquifer sediments.

The sediments in this region are of marine origin; the groundwater in strata deeper than 100 m, which is not significantly impacted by the near-surface groundwater, contains high levels of Na^+ and Cl^- originating from ancient seawater. Discrete impacts of these historical stores of sodium and chlorine can be seen in the groundwater of shallower strata. There is low potential for anion exchange, and the Cl^- ions are quickly replaced by HCO_3^- ions from groundwater flowing in from the mountainous regions. In contrast, in the case of cations, there are exchanges between sediment surface and aqueous phase; Na^+ and K^+ are more strongly partitioned to the aqueous phase than Mg^{2+}, and Ca^{2+} is more strongly absorbed to the sediment surface than Mg^{2+}. Accordingly, as the Na^+ adsorbed to sediments continues to be replaced by Ca^{2+}, the quality of the interstitial water changes from Na^+ dominant, to Mg^{2+} dominant, to Ca^{2+} dominant.

The expected partitioning of cations between aqueous and adsorbed phases is consistent with the gradient in water quality observed across the Osaka Plain.

In other words, the primary cause of geographic variability in cation content of the shallow groundwater is the exchange of cations between the marine sediments containing elements from ancient seawater and the inflowing freshwater rich in calcium. This pattern of change in groundwater quality represents a general trend in cases where calcium-bicarbonate type water from mountainous regions passes through sodium-rich marine sediments. Cation exchange at the solid–water interface of sediments that make up the aquifer has a significant impact on water quality.

4.2.5 Groundwater Research Utilizing the Geographic Variability in Water Quality of Different Aquifer Recharge Areas

Floodplain strata comprise sediments carried and deposited by rivers and streams. The chemistry of the terrestrial strata of the alluvial fan or river delta does not differ significantly from provenance rocks, although the former is subjected to chemical weathering. Therefore, different from what happens in marine sediments, the groundwater takes on the geochemical characteristics of the water that recharges it. As discussed previously, river water is generally of the $Ca^{2+} - HCO_3^-$ type, but also strongly reflects the geochemical characteristics of the watershed in which it is found. In other words, because rivers endow the water with unique qualities, these qualities can be used to trace the source and flow of groundwater.

The city of Saijo in Ehime prefecture is called the "Spring Water Capital" because of the abundant groundwater found under its populated floodplain. The water is used not only as household water but also for agriculture. Alluvial fans are developed at the base of the mountainous region, but the elevation of the floodplain as a whole is low, and its strata consist of estuarial and coastal sediments. The majority of the plain was under water at the time of the *Jomon* transgression (approximately 6,000 years ago) and the coastal area was only reclaimed beginning in the Edo period. As such, the underground strata are significantly impacted by the presence of seawater constituents. The aquifer, consisting of sand and gravel, is highly permeable, and because of the abundant inflow of water from Mt. Ishizuchi to the south, much of the groundwater has been desalinized. Thus, the majority of the groundwater is of the calcium-bicarbonate type, and, with the exception of a small area that is discussed later, is little affected by the seawater.

Figure 4.4 compares the chlorine concentrations of river water in the mountainous area near Saijo City and groundwater in the floodplain. Chlorine is used on a daily basis in households as table salt and in the manufacture of soda-lime glass, polyvinyl chloride (PVC), etc.; as a result, water in areas with considerable human activity tends to contain high levels of chlorine. Although chlorine may be found in high concentrations in some hot springs or mineral springs, in general, it only occurs in low concentration in rocks and is not subject to high rates of sediment adsorption–desorption. On the other hand, precipitation contains chlorine

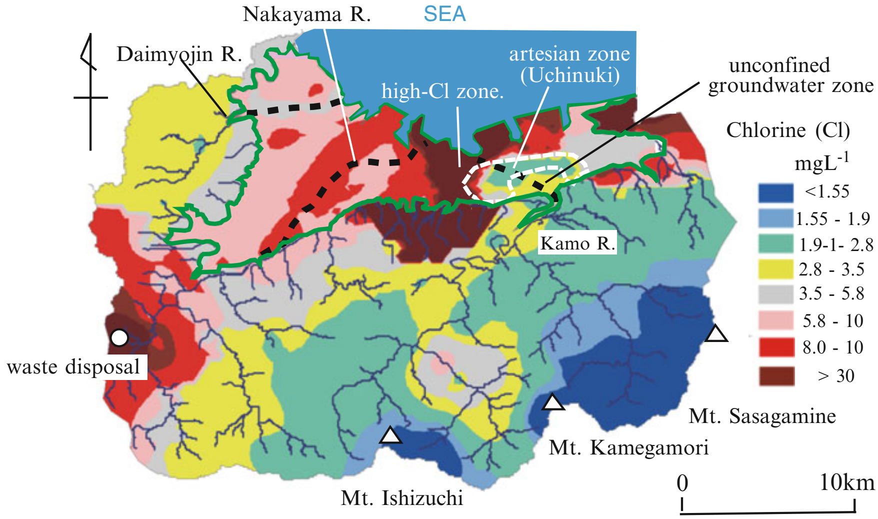


Fig. 4.4 Map of chlorine concentration in surface waters (streams and rivers) of the mountainous regions and groundwater in the floodplain area (outlined by the *bold green line*) of Saijo City, Ehime Prefecture

originating from aerosolized sea salt, with the concentration generally declining as a function of distance from the coast; as such, the freshwater exhibits geographic characteristics in terms of chlorine concentration. In other words, chlorine is a highly conserved property of water that can be used to "trace" the source of a water sample.

The chlorine concentration of the rivers flowing in the mountainous regions of Saijo City decreases with elevation. This trend is particularly striking in the high-elevation regions in the east. For this reason, rivers in the largest watersheds tend to have the lowest concentrations of chlorine. The exception is the Nakayama River, which is fed by a tributary containing high levels of chlorine as a result of the input of industrial waste and thus exhibits high concentrations of chlorine as it reaches the fan portion of its alluvial fan. Accordingly, the groundwater recharged by the Nakayama River also exhibits high concentrations of chlorine. One notable feature in Fig. 4.4 is that the outlines of the groundwater zones delineated based on chlorine concentration in the western part of Saijo City follow the paths of the rivers. For example, the Daimyojin River, as it flows in the westernmost mountainous region, is characterized as having low chlorine concentration; as the river flows through the floodplain, the groundwater near the river also exhibits low chlorine concentration. This coincidence of chlorine concentrations not only indicates that the groundwater is recharged by the Daimyojin River, but also demonstrates that chlorine concentrations can be used as traceability data to identify the recharge area.

The Kamo River, which has the largest watershed area, flows into the eastern region (Fig. 4.4). The chlorine concentration of the Kamo River is the lowest of all rivers flowing through Saijo City. Downstream, the Kamo River disappears underground to become shallow groundwater (unconfined groundwater) that flows along the surface of an aquiclude; the chlorine concentration of this groundwater is also low, and the groundwater of the artesian area located to the north (artesian aquifer), called the "uchinuki" zone, is characterized by even lower chlorine concentration. Comparing the chlorine concentrations of the rivers, it appears that the recharge area of the "uchinuki" zone is further upstream than that of the shallow groundwater. An investigation of other water quality data also confirms that the water in the artesian aquifer differs from that of the shallow groundwater recharged by the Kamo River and suggests that the recharge source is fissure water flowing through fractures in the bedrock.

Incidentally, there is a zone stretching from the center of the plain to the east, alongside the coastal zone, that exhibits extremely high chlorine concentrations. As a result of excess pumping, the groundwater pressure on the seaward side is reduced, causing seawater to infiltrate and to salinize the groundwater. However, this zone lines up almost exactly with a zone where marine clays are dominant. The fine-grained clay has a large specific surface area compared to sand and gravel, and ions, once adsorbed to the clay surface, are much less likely to desorb. Although the high groundwater chlorine concentrations are, in part, attributable to the salinization accompanying seawater infiltration, it is believed that the majority of the salinity represents constituents of ancient seawater "trapped" in the fine sediments. By mapping the distribution of chlorine, as well as many other elements, and comparing these maps with watershed water quality data, it is possible to identify the source and movement of groundwater and also capture the impacts of human activity. Water quality maps that combine surface water and groundwater data are extremely useful in investigating the groundwater environment.

4.3 Stable Isotopes

4.3.1 Stable Isotopes in Meteoric Water

The majority of elements consist of a number of stable isotopes of different atomic mass. For example, there are two stable isotopes of hydrogen (^{1}H and ^{2}H or D [deuterium = heavy water]) and three of oxygen (^{16}O, ^{17}O, and ^{18}O), although these elements overwhelmingly exist as ^{1}H and ^{16}O (Table 4.1). That is, although water is primarily made up of $H_2{}^{16}O$, it includes a very small fraction of water molecules containing heavier isotopes, such as $HD^{16}O$ and $H_2{}^{18}O$. These different molecular species are called "isotopomers," and in nature, water exhibits variability in the relative abundance of different isotopomers. In other words, the composition of stable isotopes can be used to characterize different water samples.

Table 4.1 Main stable isotopes used in groundwater studies and their relative abundances

Element	Isotope	Abundance (%)	Isotope ratio	Standard
Hydrogen	^{1}H	99.99	$^{2}H/^{1}H$	SMOW (standard mean ocean water)
	^{2}H	0.02		
Oxygen	^{16}O	99.76	$^{18}O/^{16}O$	
	^{17}O	0.04		
	^{18}O	0.20		
Carbon	^{12}C	98.90	$^{13}C/^{12}C$	Peedee belemnite (PDB)
	^{13}C	1.10		
Nitrogen	^{14}N	99.63	$^{15}N/^{14}N$	Air
	^{15}N	0.37		
Sulfur	^{32}S	95.02	$^{34}S/^{32}S$	Canyon diablo troilite (CDT)
	^{33}S	0.75		
	^{34}S	4.21		
	^{36}S	0.02		
Strontium	^{84}Sr	0.56	$^{87}Sr/^{86}Sr$	NBS987
	^{86}Sr	6.99		
	^{87}Sr	9.86		
	^{88}Sr	82.59		

In general, the relative abundance of a stable isotope is expressed as the ratio of the second most common isotope to the most common isotope. As these fractions are very small and difficult to work with, they are reported as relative differences (δ values) to a known reference in parts per thousand, calculated as follows:

$$\delta\mathrm{D} = \left[\left(\mathrm{D}/^{1}\mathrm{H}\right)_{\mathrm{sample}} \Big/ \left(\mathrm{D}/^{1}\mathrm{H}\right)_{\mathrm{reference}} - 1\right] \times 1000 \tag{4.4}$$

$$\delta^{18}\mathrm{O} = \left[\left(^{18}\mathrm{O}/^{16}\mathrm{O}\right)_{\mathrm{sample}} \Big/ \left(^{18}\mathrm{O}/^{16}\mathrm{O}\right)_{\mathrm{reference}} - 1\right] \times 1000 \tag{4.5}$$

Although the isotope ratios of both hydrogen and oxygen in ocean water vary slightly across the ocean environment, based on the fact that oceans hold 97.5 % of the earth's water and play an important role in the hydrologic cycle, a standardized average ocean water (Standard Mean Ocean Water) is used as the reference standard for both elements. In contrast, although the δD and $\delta^{18}O$ values of wet deposition vary significantly in both time and space, there is a linear relationship between the two:

$$\delta\mathrm{D} = a\delta^{18}\mathrm{O} + b \tag{4.6}$$

In hydrology, water originating from wet deposition is referred to as meteoric water. Based on an analysis of approximately 400 samples of meteoric water (wet deposition, rivers, lakes, and marshes) from around the world, Craig (1961) determined the relationship to be

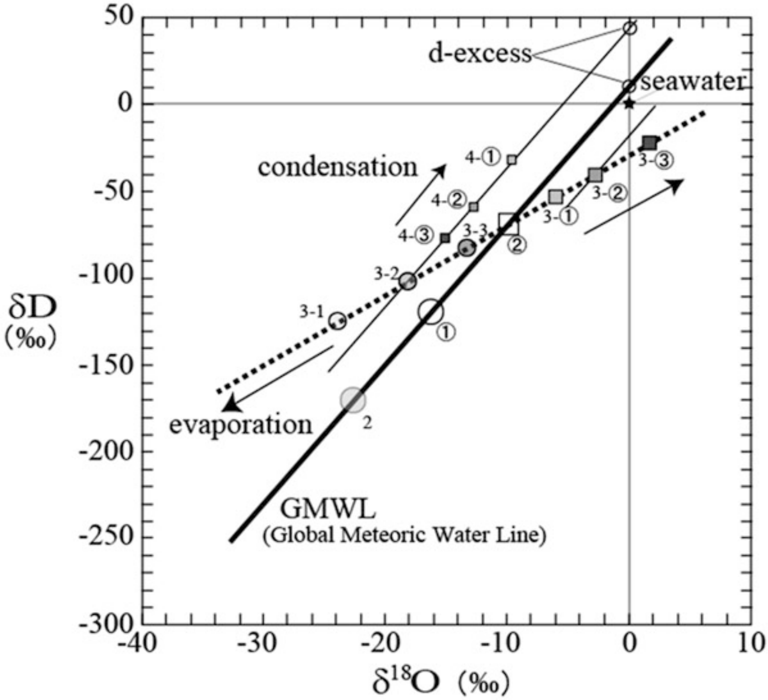


Fig. 4.5 Relationship between stable isotopes of hydrogen and oxygen in meteoric water. When the terrestrial water ② arising from condensation of water vapor ① evaporates, depending on the degree of evaporation, it results in water vapor (3-1, 3-2, or 3-3). The remaining terrestrial water is concentrated and becomes water with low *d*-values (3-①, 3-②, 3-③). Water arising from condensation of water vapor 3-2 (generated by evaporation of terrestrial water) has different δD and δ^{18}O values (4-①, 4-②, 4-③) depending on the amount of condensation. Furthermore, each of these waters exhibits high *d*-values and the line fitted to these points plotted on the chart has a slope of approximately 8

$$\delta D = 8\delta^{18}O + 10 \tag{4.7}$$

as outlined in Fig. 4.5.

Presently, wet deposition is being collected around the world by groups such as the International Atomic Energy Agency, and the isotope ratios of hydrogen and oxygen in tens of thousands of samples are being analyzed. As the number of measurements increases, the geographic and seasonal variability in hydrogen and oxygen isotope ratios in meteoric water becomes apparent. However, the linear relationship discovered by Craig [Eq. (4.7)] remains valid and is referred to as the global meteoric water line (GMWL).

The variation in isotopic ratios of hydrogen and oxygen in meteoric water results from isotopic fractionation during evaporation and condensation within the hydrologic cycle. Compared to water vapor, liquid water consistently contains higher fractions of the heavier stable isotopes (^{2}H and ^{18}O). Because raindrops are generated in clouds saturated with water vapor, the isotopic ratios of the two are generally in equilibrium. Although the equilibrium constant of hydrogen between liquid water and water vapor is larger than that of oxygen, the two exhibit similar dependencies on temperature. The variable *a* in Eq. (6) can be expressed as a function of the partition coefficients of hydrogen and oxygen $[(\alpha_{hydrogen} - 1)/(\alpha_{oxygen} - 1)]$; at temperatures above 0 °C, the value, regardless of temperature, is approximately 8. This value is reflected in the slope of the GMWL.

The variability in $\delta^{18}O$ value (or δD value) in meteoric water can be described by the following Rayleigh distillation model:

$$\delta = (\delta_0 + 1000)f^{\alpha - 1} - 1000 \tag{4.8}$$

where δ_0 is the initial $\delta^{18}O$ (or δD) value of water vapor, α is the isotopic fractionation coefficient of oxygen (α_{oxygen}) [or of hydrogen ($\alpha_{hydrogen}$)] between liquid water and water vapor, and *f* expresses the progress of the reaction and is the ratio of the remaining water vapor to the initial vapor. As raindrops are removed from the water vapor (i.e., as precipitation increases), the $\delta^{18}O$ and δD values of the vapor decrease; however, because isotopic fractionation is an equilibrium process, the $\delta^{18}O$ and δD values of the precipitation also change in a similar fashion (Fig. 4.5). The GMWL can be explained by a Rayleigh distillation model whereby precipitation is continuously removed from rain clouds formed from the large amounts of water vapor generated by the evaporation of oceans in lower latitudes.

The evapotranspiration that gives rise to water vapor is affected by factors such as temperature and humidity, but the advection and diffusion of water vapor is a nonequilibrium process. The environment in which the water vapor is generated is expressed by the variable *b* in Eq. (4.6), and is referred to as the deuterium excess (here, *d*-value). In the case of meteoric water caused by rapid evaporation, the *d*-value is large; when evaporation is slow, the *d*-value becomes smaller. It can also be said that the *d*-value characterizes the air mass that gives rise to the meteoric water. In Japan, the *d*-value of the snow and rain falling on the Japan seaside during winter is high (20–35) while that of the precipitation caused by the Ogasawara air mass in the summer is low (0–15).

4.3.2 Geographic Characteristics of Hydrogen and Oxygen Stable Isotopic Ratios

As discussed, it is known that the δD and $\delta^{18}O$ values of meteoric water vary substantially in time and in space. Geographic variability is particularly large, with decreasing δD and $\delta^{18}O$ associated with increasing latitude (Fig. 4.6). This “latitude effect” is especially evident in North America. At the same time, δD and $\delta^{18}O$ values of water samples collected at the same latitude are observed to also be

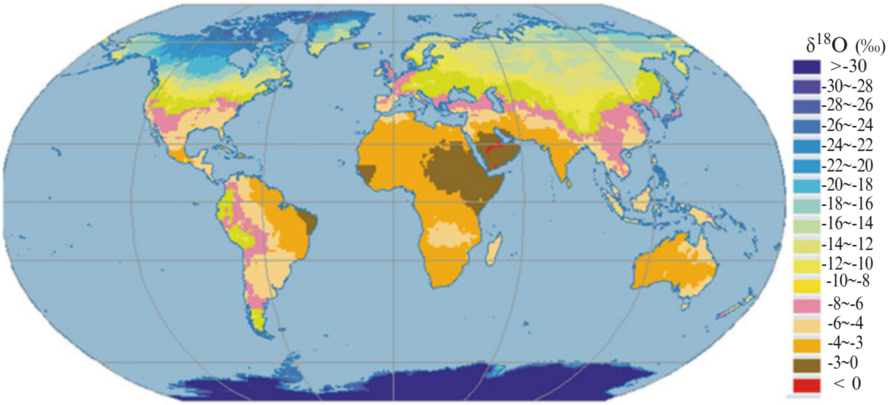


Fig. 4.6 International Atomic Energy Agency (IAEA) interpolated map of oxygen isotope ratios in meteoric water (http://nds121.iaea.org/wiser/gui/map.php)

elevation dependent, with lower values being associated with higher-elevation areas such as the Rocky, Sierra Nevada, or Appalachian Mountains: this is referred to as the "altitude effect" or "elevation" effect."

Temperature decreases with increasing latitude or elevation and thus exhibits a strong correlation with the isotopic ratios of hydrogen and oxygen. This effect, then, is called the "temperature effect." The temperature effect is particularly evident in the mid- to high latitudes; in regions where the monthly average temperature is 15 °C or less, the $\delta^{18}O$ value of precipitation decreases between 0.2 ‰ and 0.6 ‰ (global average = 0.53 ‰) for every 1 °C decrease in temperature. The temperature effect varies not only with location, but also with time (e.g., season). In winter, when temperatures are low, the magnitude of the decrease in $\delta^{18}O$ value is generally large. In addition, δD and $\delta^{18}O$ values vary with precipitation volume, which is called the "amount effect." In general, δ values tend to decrease with increasing precipitation; in contrast to the temperature effect, the amount effect is particularly evident in lower latitudes (below 40° N or S) and is not seen at higher latitudes. For every 100-mm increase in precipitation, $\delta^{18}O$ values of meteoric water decrease by 1.0–6.0 ‰.

With respect to evapotranspiration, which is governed by water vapor dynamics, the a-value of water in Eq. (4.6) that has evaporated (from lakes and marshes), as well as that of the remaining water, is between 3 and 5, clearly smaller than the slope of the GMWL (≈8). Consequently, d-values of meteoric water from clouds that form as a result of terrestrial evapotranspiration are high whereas those of the lake or marsh water are low (Fig. 4.5). In this manner, the temporal and spatial fluctuations of the isotope ratios in precipitation are intimately related with the factors that cause phase changes of water and, as such, are used for validation of atmospheric hydrologic models (Ichiyanagi 2007; Yoshimura et al. 2009).

During subterranean flow, groundwater does not evaporate; although it may react with the surrounding rock (sediment), no new water is added. As a result,

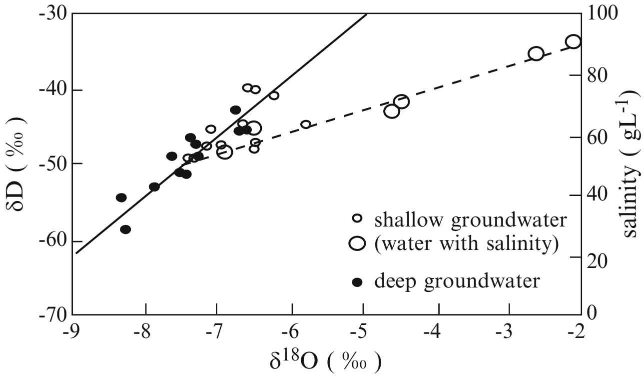


Fig. 4.7 Relationship between δD and $\delta^{18}O$ values of groundwater in Algeria. *Solid line* represents a local meteoric water line. (Modified from Mazor 2004)

the geographic and seasonal characteristics of the δD values, $\delta^{18}O$ values, and *d*-values of the precipitation falling in a given watershed are excellent tracers for identifying groundwater recharge areas (Mazor 2004). The isotopic ratios of groundwater, even in the same sedimentary basin, generally vary with location and depth. For example, in Manitoba Province in central Canada, groundwater δD and $\delta^{18}O$ values lie directly on the regional precipitation line ($\delta D = 8.1 \cdot \delta^{18}O + 11$), indicating that its source is meteoric water.

On the other hand, in the Algerian example presented in Fig. 4.7, the δD and $\delta^{18}O$ values of the deep groundwater differ from those of the shallow groundwater. Although it is possible that the deep groundwater is recharged by a source at a higher elevation, based on its old age (determined by radiocarbon dating) it is believed that it represents meteoric water recharged under different climatic conditions. In contrast, based on the fact that, according to the $\delta D - \delta^{18}O$ map, the *a*-values of the shallow groundwater are low (≈3), it can be determined that the source of the shallow groundwater has undergone evaporation. In fact, the shallow groundwater with the highest δD and $\delta^{18}O$ values exhibits the highest salt content, which is consistent with the isotope ratio data. In the case of groundwater that is recharged by lake water that has undergone significant evaporation, by combining *d*-value data with other measures of water quality, it is possible to identify the recharge area with even greater accuracy.

Although the "elevation effect" on meteoric water varies by region, because both the $\delta^{18}O$ and δD values decrease with elevation (0.2–0.6 and 1.5–5 ‰/100 m, respectively), they are useful in delineating groundwater recharge areas. In the case of surface water, the elevation effect becomes clearer when the average elevation is used. It is also possible to use the temperature effect, amount effect, or temporal fluctuation in isotope ratios of precipitation to delineate groundwater recharge areas (Bortolami et al. 1970). In any case, when the precipitation and surface water exhibit temporally and spatially distinct characteristics, the data can be used to determine the source of a groundwater (i.e., traceability).

4.3.3 Sulfur and Nitrogen Isotopes: Oxidation–Reduction

Nitrate, bicarbonate, and sulfate anions in groundwater all contain oxygen. Dissolved oxygen in groundwater is consumed through respiration by living organisms. When the dissolved oxygen is used up, nitrate ions are used first to be reduced. This process, called denitrification, generates nitrogen gas according to the following reaction:

$$2\mathrm{NO}_3^- + 5\mathrm{H}_2 \rightarrow \mathrm{N}_2 + 4\mathrm{H}_2\mathrm{O} + 2\mathrm{OH}^- \tag{4.9}$$

When the nitrate is used up, oxidized forms of iron and manganese are reduced, resulting in an increase in the aqueous concentration of divalent iron and manganese. As reductive processes continue, sulfate ions are reduced as follows, generating hydrogen sulfide gas:

$$2\mathrm{CH}_2\mathrm{O} + \mathrm{SO}_4^{2-} = \mathrm{H}_2\mathrm{S} + 2\mathrm{HCO}_3^- \tag{4.10}$$

These reductive reactions are carried out by microorganisms. There are two stable isotopes of nitrogen and four stable isotopes of sulfur (Table 4.1), the isotopic ratios of which are expressed as follows:

$$\delta^{15}\mathrm{N} = \left[\left({}^{15}\mathrm{N}/{}^{14}\mathrm{N}\right)_{\mathrm{sample}} \Big/ \left({}^{15}\mathrm{N}/{}^{14}\mathrm{N}\right)_{\mathrm{reference}} - 1\right] \times 1000 \tag{4.11}$$

$$\delta^{34}\mathrm{S} = \left[\left({}^{34}\mathrm{S}/{}^{32}\mathrm{S}\right)_{\mathrm{sample}} \Big/ \left({}^{34}\mathrm{S}/{}^{32}\mathrm{S}\right)_{\mathrm{reference}} - 1\right] \times 1000 \tag{4.12}$$

The concentrations of nitrate and sulfate ions decrease with progression of reductive processes. Because the lighter isotopes of both nitrogen and sulfur are partitioned to nitrogen and hydrogen sulfide gas, the δ^{15}N and δ^{34}S values of the groundwater increase with the advancement of reduction. Furthermore, the δ^{18}O value of the oxygen contained in both nitrate and sulfate increases (Lehmann et al. 2002; Hosono et al. 2011; Heffernan et al. 2012).

In the aforementioned Takatsuki example, a negative correlation can be seen between the dissolved ion concentration and stable isotope ratio (Yamanaka et al. 2007). This relationship can be reproduced by a model describing the sequential reduction of groundwater sulfate ions through microbial activity. When sulfate undergoes reduction in a closed system, the δ^{34}S value can be described using the same Rayleigh distillation model as in Eq. (4.8):

$$\delta^{34}\mathrm{S} = \left(\delta^{34}\mathrm{S}_0 + 1000\right) f^{\alpha-1} - 1000 \tag{4.13}$$

where $\delta^{34}\mathrm{S}_0$ is the initial δ^{34}S value of the sulfate ions, α is the partition coefficient between the sulfate ions and hydrogen sulfide gas, and f is the ratio of the instantaneous sulfate concentration to the initial sulfate concentration. Let us take

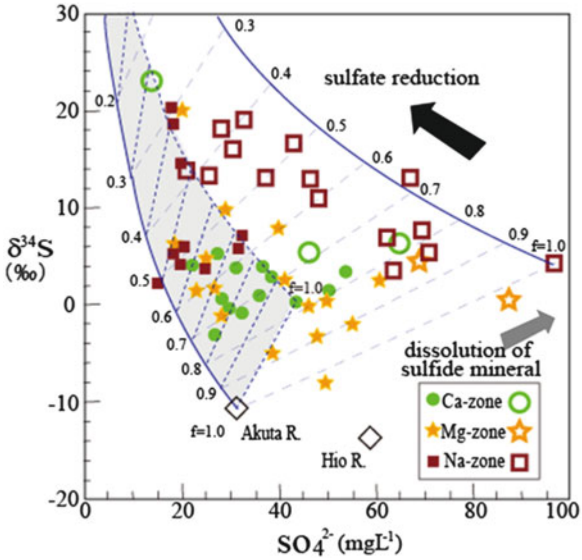


Fig. 4.8 Relationship between the sulfur stable isotope ratio in aqueous sulfate and sulfate concentrations in groundwater of northeastern Osaka City. (Modified from Yamanaka et al. 2005)

the case of the shallow groundwater in Fig. 4.8 as an example. If we assume the Hio River in the mountainous region to be the groundwater source (i.e., the initial condition in the model above), and if we estimate α to be 20, the model yields a reasonably good fit to the actual data, which demonstrates that the observed correlation can be explained by such a model.

There is a fault running through the same region, and the groundwater to the west of the fault is mapped as being completely different from that to the east. The marine strata contain sulfide minerals (primarily pyrites) that release high concentrations of sulfate ions upon oxidative dissolution. The sulfate concentration and δ^{34}S value of the groundwater to the west are in the range that would be expected if groundwater from the mountainous regions were to mix with sulfate ions originating from the pyrites and then undergo reduction. Denitrification results in similar changes wherein the δ^{15}N value of the groundwater increases with progression of denitrification.

The mixing of two types of groundwater differing in quality can explain the relationship observed in Fig. 4.8, but this does not explain the existence of water with high concentrations of SO_4^{2-} but low δ^{34}S values. Recently, it has become possible to simultaneously analyze the isotope ratio of the oxygen contained in sulfate and nitrate ions, enabling researchers to explain the processes involved in the formation of groundwater quality in greater detail. Simultaneous analysis of the concentrations and stable isotope ratios of both ions (nitrate and sulfate) should prove useful in separating out the impacts of human and microbial activity during the process of groundwater flow.

4.3.4 Strontium Stable Isotopes: A Cation Traceability Index

Although four stable isotopes of strontium exist (Table 4.1), ^{87}Sr also is generated by the radioactive decay of ^{87}Rb. Based on the fact that this decay is extremely slow and that strontium is found in rocks and minerals, albeit in very small quantities, the Rb–Sr method is widely used for geologic dating of rocks. It has been demonstrated in numerous studies that rocks retain an ^{87}Sr/^{86}Sr ratio that is characteristic of the environment in which they were formed. Sr and Ca are both alkaline earth elements, and as a consequence, their geochemical behaviors are very similar. In general, the Sr concentration in water is approximately 1 % that of the dominant element Ca, and the concentrations of the two elements are strongly correlated. As described, Ca-containing minerals are readily weathered, resulting in relatively high concentrations of dissolved calcium in water (generally several tens or hundreds of μg l^{-1}) whose ^{87}Sr/^{86}Sr ratio closely reflects the geologic environment.

For example, limestone retains the ^{87}Sr/^{86}Sr ratio of the seawater at the time that it was formed, and although the ratio may vary depending on the limestone's geologic age, it generally falls within the range of 0.7080 ± 0.0012. Groundwater that flows through limestone caverns exhibits very similar ^{87}Sr/^{86}Sr ratios to that of the limestone. In contrast, groundwater that flows through lava or products of volcanic eruptions exhibits the low ^{87}Sr/^{86}Sr ratios characteristic of volcanoes. For example, the rocks of Mt. Fuji characteristically exhibit ^{87}Sr/^{86}Sr ratios (≈0.703) that are low even by volcano standards, and the spring water in the same area exhibits similarly low ratios.

Similarly, in the strata that constitute floodplains, the ^{87}Sr/^{86}Sr ratio of the groundwater varies with that of different aquifers. This tendency is particularly strong in the case of terrestrial strata. Alternatively, this means that it is also possible to recreate the geologic characteristics of an aquifer based on the Sr isotope ratio of the groundwater. The temporal variability in the Sr isotope ratio of groundwater is especially low, which indicates that the Sr is strongly buffered by the surrounding rock and aquifer. In the case of marine strata, the stable isotope ratio of the Sr adsorbed to the surface of the sediment, which originates from ancient seawater, generally differs from that of the Sr contained in the minerals that comprise the sediment. When freshwater originating in mountainous regions flows into such marine strata to become groundwater, it undergoes ion exchange with the Sr that has been preserved by adsorption to the sediment surface. It is possible to identify the source of Ca, which is the dominant component of freshwater, by utilizing the stable isotope ratio of Sr as a proxy for Ca.

In the Osaka Plain example, we can observe an overall positive correlation between Ca and Sr (Fig. 4.9). However, the correlation between Sr and Ca differs in the groundwater of the Ca-rich zone, which is strongly influenced by the surface water, the Mg-rich zone, which is strongly affected by ancient seawater, and the Na-rich zone, suggesting that the origin of the two elements is different in each zone. The Sr concentration does not vary much from the Ca-zone to the Mg-zone, but the ^{87}Sr/^{86}Sr ratio gradually approaches that of seawater (≈0.7092), indicating

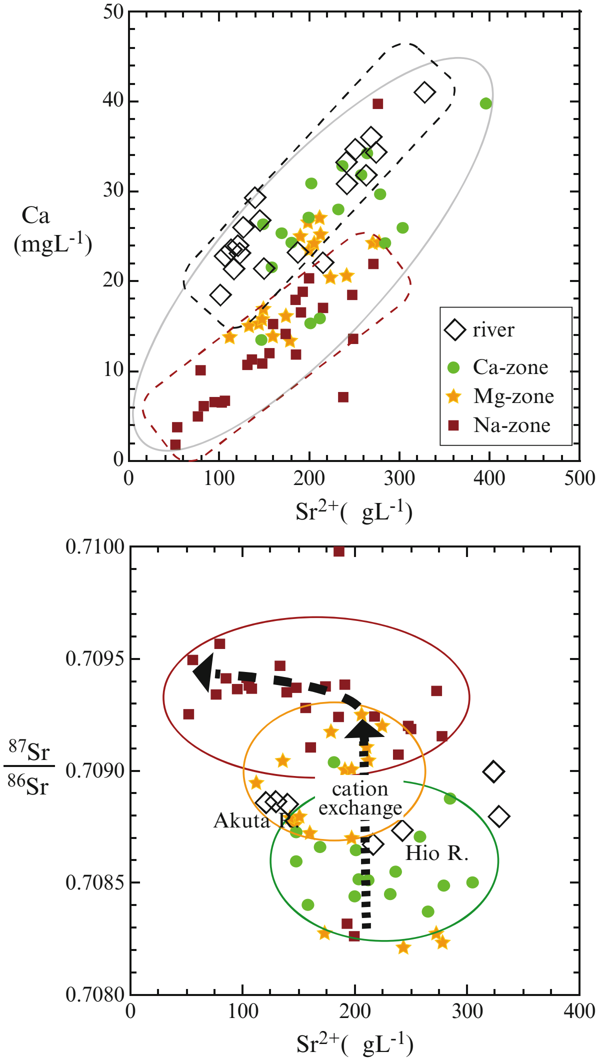


Fig. 4.9 *Upper panel*: Relationship between Sr and Ca concentrations in groundwater and river water in northeastern Osaka City. *Lower panel*: Relationship between Sr concentration and Sr stable isotope ratio in the same water. (Modified from Yamanaka et al. 2005)

that the Sr from the recharge area is exchanged with the Sr originating from ancient seawater. In contrast, from the Mg-zone to the Na-zone, although the $^{87}Sr/^{86}Sr$ ratio remains approximately the same as that of seawater, the concentration of Sr declines. It is believed that the Sr originating from the ancient seawater that was released into solution by cation exchange, then, is exchanged for the Na that is abundantly adsorbed to the marine sediment. A similar process accounts for the spatial variability of Ca, with which the Sr is strongly correlated. The $^{87}Sr/^{86}Sr$ ratio of groundwater has been shown to be distinctly different from that of the sediment. Furthermore, in experiments in which mountain water is passed through a glass column packed with aquifer clay, it has been possible to reproduce the quality and Sr isotope ratio of the groundwater observed in the field (Yamanaka et al. 2005). These results indicate that cation exchange is an important factor in the variability in quality of groundwater.

4.4 Summary

Water, with its constituent elements and stable isotopes, contains "traceability information" related to hydrologic and material cycles. The stable isotope ratios of hydrogen and oxygen (δD values, $\delta^{18}O$ values, and *d*-values) contained in meteoric water and surface waters originating from meteoric water vary both in space and in time; by characterizing and comparing this variability, it becomes possible to delineate the recharge areas of groundwater.

The various constituents of groundwater can be classified based on their origin, that is, meteoric water, watershed geology, human activity, etc. In humid regions such as Japan, not only the precipitation but also the soil water is acidic. This acidic water is neutralized as it dissolves minerals in rock and soil, eluting cations. This neutralization reaction generally involves carbonic acid and results in the water taking on qualities of the minerals that it weathers. Because calcium minerals are easily weathered, freshwater is generally of the $Ca\text{-}HCO_3$ type; however, in rock, such as granite, that is rich in sodium or in regions that are significantly impacted by sea salt aerosols, the freshwater is of the $Na\text{-}HCO_3$ or the Na-Cl type.

The input of human-derived constituents in addition to those introduced through the natural processes discussed results in terrestrial water having distinct geographic characteristics. The groundwater flowing through a floodplain aquifer strongly reflects the quality of the water that recharges it. However, in the case of marine sediments, the interstitial water contains elements derived from seawater and sulfide-bearing minerals such as pyrites, which significantly alter the quality of the incoming recharge water. Further, given the slow movement of groundwater, the consumption of dissolved oxygen by microbial activity, etc. creates a reducing environment. In such a reducing environment, the decreasing concentration of dissolved oxygen, nitrate, and, subsequently, sulfate is accompanied by an increase in aqueous concentrations of iron, manganese, and heavy metals such as arsenic that are strongly adsorbed to iron oxides.

Nitrogen and sulfur are contained in household wastewater and fertilizer, for example, and are introduced into the water environment as a result of human activity; as such, the stable isotope ratios of nitrate and sulfate can be useful indicators of human impact. The stable isotope ratios of all elements are known to increase with progression of reductive processes. Thus, simultaneous analysis of stable isotope ratios and elemental content enables researchers to recreate the chemical environment encountered by groundwater as it moves through a system.

Presently, methods exist to analyze more than 50 elements. By also analyzing the elements eluted in a reducing environment, it is possible to more accurately determine the role and contribution of natural and human factors in the groundwater environment.

There are stable isotopes of numerous cations. In particular, because the geochemical behavior of Sr is similar to that of Ca, the dominant constituent of water, the $^{87}Sr/^{86}Sr$ ratio of groundwater can be used to recreate the geologic environment of an aquifer or bedrock. Because the $^{87}Sr/^{86}Sr$ ratio of seawater is constant globally and temporally, it can be used to determine the extent of salinization. Currently, it is becoming possible to analyze the stable isotope ratios of various metals. The constituents and stable isotope ratios of water are now poised to be broadly used in groundwater management as “traceability data” for determining the source and flow of groundwater.

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Chapter 5
Ecohydrological Assessments on Nitrogen Behavior in the Headwater Wetland

Changyuan Tang

Abstract From the aspect of ecohydrology, a wetland is the discharge area for the groundwater system and supports habitats of high biodiversity. A typical headwater wetland in Chiba, Japan, was chosen to investigate nitrate behavior in the wetland through which the groundwater flowed.

As one of the predominant nitrogen species, nitrate was removed completely from the groundwater in three stages associated with activities of denitrifying bacteria in the aquifer where N_2O was produced as an intermediate product and dissolved in the groundwater. It was found that most of the dissolved N_2O was transformed to dissolved N_2 in the groundwater because strong reduction conditions prevailed in the wetland aquifer. As a result, all nitrate entering the wetland was eventually transformed to nitrogen gas and N_2O that was released to atmosphere at the seepage surface toward which the groundwater flowed. Also, the groundwater flow path was the vital channel providing nitrate for the denitrifying bacteria community in the wetland. When the high nitrate of groundwater flowed into the wetland, the microbial denitrifier community extended with DO depletion and the sharp decrease of ORP. The chemoautotrophic denitrification caused a decline of NO_3^- and increase of SO_4^{2-} in the groundwater at the same time. The groundwater generated conditions suitable for the microbial community to survive in the aquifer and developed the reduction sequence of DO, nitrate, and sulfate along the pathway in the wetland. Therefore, one of the ecohydrological functions of the headwater wetland should be considered as a transformer between the biogeochemical source and sink for nitrate to maintain the ecosystem services in the watershed.

Keywords Denitrification • Denitrifying bacteria • Headwater wetland • N_2O • Nitrate

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5.1 Introduction

As one of the most sensitive areas in the world, wetlands are transitional buffers between terrestrial and open-water aquatic ecosystems that have been investigated for a long time by biologists and ecologists because of their biological diversity. Affected by climate, basin geomorphology, and hydrology, the wetland becomes an important area for interaction between groundwater and surface water. For example, a wetland at a lowland or headwater is one kind of wetland connected with a slope (Brinson 1993). It is usually found between uplands and aquatic systems as the discharge area where groundwater flows out as springs and/or seepages that become the source of rivers in the watershed. As a result, wetland hydrology is the driving force that determines not only soil development, the assemblage of plants and animals that inhabit the site, but also the type and intensity of biochemical processes. The hydrology of a wetland directly modifies and changes its physicochemical environment, such as pH, oxygen availability, and nutrients as well as toxicity. Modifications of the physicochemical environment, in turn, have a direct impact on the biota in the wetland, which is the key for the wetland functioning ecologically.

Geomorphologically, water and chemicals inputs from the upland are the major source of nutrients to a headwater wetland where the aquifer consists of hydric soils. The soils are formed under saturation conditions because the seepage or spring develops an anaerobic environment all year round. The soils may be composed mostly of mineral constituents (sand, silt, clay), and contain large amounts of organic matter because anaerobic conditions slow or inhibit decomposition of organic matter. Usually the upland and wetland are linked by the slope from the surface of which soil and minerals can move to the wetland.

Wetlands can provide to us with many ecological services upon which we are and have been relying. One of the most essential benefits of wetland systems is their capability for improving water quality, thus being a habitat that has a very important ecological function. Bordering agricultural lands where the groundwater commonly contains high concentration of nitrate, wetlands become the essential buffer zones for agricultural ecosystem because of their ability for freshening the groundwater (Dørge 1994; Blackwell et al. 1999; Tang et al. 2004; Zaman et al. 2008). As a result, the wetland plays a great role in the natural attenuation of chemicals, such as nitrate in the watershed.

Nitrogen transformations in a wetland are a complex assortment of processes mediated by microbes and strongly influenced by the redox status of the wetland soil (Faulkner and Richardson 1989). Denitrification is an anaerobic process in which the nitrate acts as an electron acceptor and organic carbon as the electron donor in a microbially mediated reduction environment (Chapelle 1993). As a result, nitrate and nitrite are reduced by bacteria, yielding nitrous oxide and nitric oxide, which may be further reduced to nitrogen gas (Knowles 1982). Nitrous oxide (N_2O), one of the so-called greenhouse gases, can be produced in subsoil, groundwater, streams, rivers, and estuaries. Soils and aquifers are the main sources of N_2O emissions, contributing 70 % of the global N_2O emission (Deurer et al. 2008). Therefore, any attempt to assess the wetland function of denitrification should understand the conditions necessary for minimizing N_2O production.

Most studies on denitrification in wetlands have inferred nitrate losses based on measurements of its decrease across the wetlands, evaluated by nitrogen isotope fraction or multi-isotope tracers (e.g., Wada et al. 1975; Bottcher et al. 1990; Aravena and Robertson 1998; Tang et al. 2004; Søvik and Mørkved 2008) or emissions of N_2O or N_2 during this process (Wilson et al. 1990; Blicher-Mathiesen et al. 1998; Mookherji et al. 2003), but few studies have explained the nitrate removal in wetland ecosystems by the comprehensive methods of microbiology and hydrogeochemistry (Hedin et al. 1998).

To understand how the functions of a wetland work, we should integrate knowledge from hydrology, biology, geomorphology, and ecology. It is important to assess the processes of nitrogen removal from groundwater in the wetland through which it flows, to recover the huge amounts of water as a raw water resource, and to protect rivers from point and non-point pollution: this becomes the underlying concept objective of this study. The purposes of this study are (1) to make clear the migration processes of nitrate in terms of geochemistry in the wetland, (2) to assess the behavior of nitrous oxide (N_2O) as the product of denitrification in the groundwater, (3) to identify the spatial distribution of denitrifying bacteria in the topsoil and shallow groundwater, and (4) to discuss the effects of hydrological functions on nitrogen in the groundwater of a headwater wetland system.

5.2 Site Description

Geomorphologically, the Ochi experimental area, Chiba, Japan (35°31′55″N, 140°14′53″E) is located at a typical headwater wetland, as the discharge area of groundwater from the upland where both rainfall and chemicals enter the farmlands with rotated planting of crops and vegetables (Fig. 5.1, left). The elevation of the upland is more than 70 m above sea level. Elevation difference between the upland and the experimental wetland ranges about 22–24 m. Covered by broad-leaved woods and conifers with some understory, the slope is transitional in terms of spatial arrangement to link the upland and the experimental wetland. To monitor the changes of wetland groundwater, seven sets of piezometers have been set at several depths from 0.5 to 4 m for each site along the cross section A–A′ where groundwater flows through the wetland (Fig. 5.1, right). A spring downstream discharges all year round and has been the only source for irrigating the paddy fields adjacent to the study wetland for several decades.

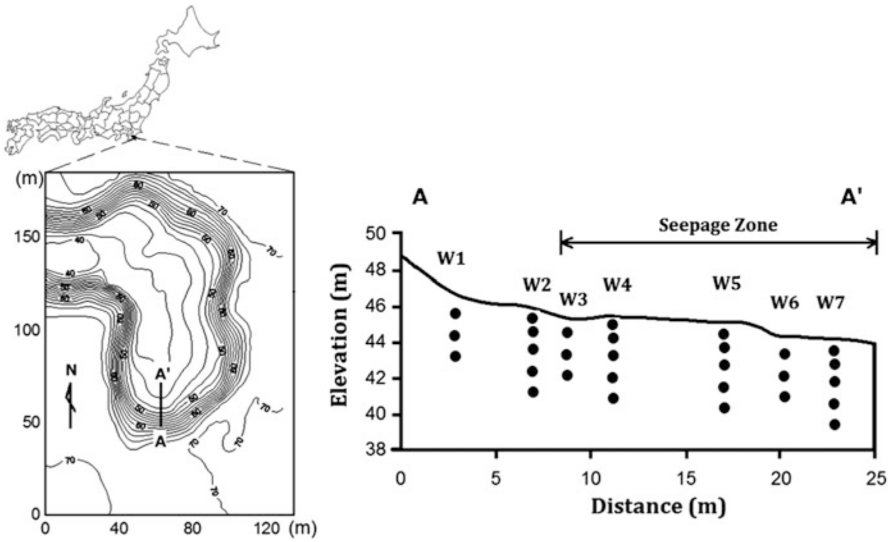

Fig. 5.1 Map of the study area (*left*) and sampling points along the cross section A–A′ (*right*)

5.3 Methods

To understand the variations of hydraulic potentials and their chemical characteristics, electrical conductivity (EC), pH, oxidation–reduction potential (ORP), and dissolved oxygen (DO) of groundwater were measured in situ from piezometers at W1 to W7 every 2 months from 2004 to 2007 (Fig. 5.1). Also, groundwater was sampled from the piezometers to understand the denitrification processes associated with major ions and denitrifying bacteria. Soil samples were taken at W1 to W7 in three depths (0–5, 5–10, 10–20 cm) with two samples of 100 g fresh soil for each site. Soil samples for bacterium incubation were taken by sterilized tools and then put into plastic bags pre-treated with 75 % ethanol. All samples were stored at 4 °C and brought to the laboratory for analysis within 24 h. Soil water for major ion analysis was extracted by centrifuge (Kokusan, Japan) with a rotating speed of 3,000 revolutions per minute for 3 min, which is about a pF value of 2.3. It was easy to obtain the water in this way because of the high water content of wetland soils. All water samples and soil waters were filtered (0.45 μm) before analyzing major ions by ion chromatography (Shimadzu CDD-6A and CDD-10Avp, ± 5–10 %).

N_2O gas was taken from the surface with the closed chamber method, using a rectangular plastic chamber (35 cm wide × 20 cm long × 35 cm high). Two holes were made separately on two sides of the chamber to accommodate rubber septa, through which a needle could be inserted on the one side of the chamber to withdraw gas samples, and a thermometer on the other side to measure the chamber air temperature. The chamber was set on the water-filled turf at the land surface, and then four gas samples were withdrawn from the chamber in 30 min at an interval of 10 min.

To measure N_2O dissolved in groundwater, a water sample of about 10 ml was injected into an evacuated 20-ml vial using a sterile, plastic gas-tight syringe immediately after the groundwater was taken from each piezomenter. The vials were injected with a few drops of sterilant to control further N_2O production, then filled with pure N_2, and vigorously shaken by hand for 1 min on site. After samples were delivered to a laboratory, the exact amount of each water sample was determined by weighing the vial before and after sampling. The gas samples and the headspaces of groundwater samples were analyzed for N_2O concentration by gas chromatography (GC14B; Shimadzu) equipped with an electron capture detector (ECD). The measurements of N_2O concentrations were completed within 24 h after collecting the samples. Gas fluxes were calculated, and cumulative N_2O emissions were estimated from the individual fluxes and the time for the measurements. N_2O dissolved in the water was calculated with the headspace N_2O concentration and Henry's law.

Counts of denitrifying bacteria were based on NO_3^-- or NO_2^--free test tubes that contained substrate and were inoculated with dilutions of soil or groundwater. The dilutions of soil or water (surface water and groundwater) were prepared by vigorously shaking 10 g fresh soil or 10 ml water sample in 90 ml saline solution to which a drop of Tween 80 (France) had been added. Then, a series of tenfold dilutions at seven levels was used to inoculate five tubes for each dilution level. After 2 weeks of anaerobic incubation at 25 °C in a light-shielded thermostatic chamber, the tubes with denitrifiers were identified by depletion of both NO_3^- and NO_2^-, as evidenced by a negative spot test with diphenylamine. Denitrifying bacteria were counted as the number per gram dry soil (cells/g dry soil) for soil and the number per milliliter for water (cells/ml) by the most probable number (MPN) method (Lorch et al. 1995). For the sake of comparison, the number of denitrifying bacteria in the soil was converted to cells/ml by multiplying the volumetric content of soil water.

To ensure no dolomite, siderite, or biogenic calcite was present in the soil, it was pre-treated at room temperature with a 3 mol/l HCl solution: 500 mg air-dried soil sieved by 80 mesh was used to analyze the C:N ratio at 1,800 °C by the dry combustion method (CN CORDER MT-700; Yanaco, Japan), and the total C was considered as organic carbon in the study area (Nelson and Sommers 1982).

5.4 Results

5.4.1 Dynamics of Groundwater in the Study Site

From the point of view of hydrogeomorphology, the upland (recharge area) and the experimental wetland (discharge area) are linked by the slope of the valley to form the field for groundwater that can flow from the upland to the valley. Wetland vegetation was present during the growing season from March to November every year. Shrubs were predominant from W1 to W3 with some understory.

Table 5.1 Hydraulic conductivity, porosity, average diameters, and other physical parameters of soil and aquifer at sites W4 and W6

Site	Depth (cm)	Hydraulic conductivity (cm/s)	Porosity (%)	Clay (%)	Silt (%)	Fine sand (%)	Sand (%)	Average diameter (mm)
W4	100	1.19×10^{-5}	60.1	15.4	35.1	24.9	24.7	0.30
W4	200	2.97×10^{-5}	66.6	17.4	32.6	25.5	24.5	0.30
W4	300	1.24×10^{-4}	58.1	23.7	25.6	33.7	17.1	0.23
W6	100	3.54×10^{-3}	54.9	13.7	19.8	43.3	23.2	0.31
W6	200	4.75×10^{-4}	56.2	23.1	19.1	33.3	24.5	0.31
W6	300	2.29×10^{-4}	50.7	24.6	26.4	24.1	24.9	0.30

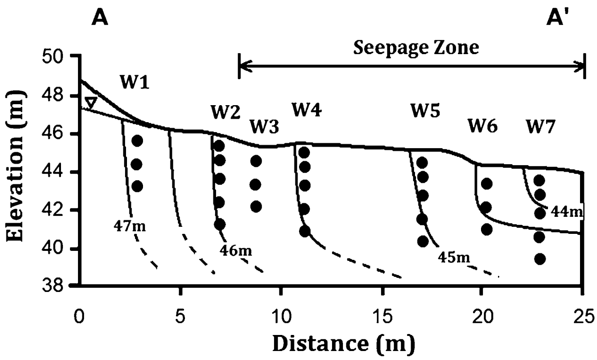

Fig. 5.2 Distribution of hydraulic potential in groundwater along A–A′ (based on the data from 2004 to 2007)

Reeds, Japanese pampas grass, tall goldenrod, etc. were seen in the area from W4 to W7. Groundwater is the most dynamic factor for water movement and migration of chemicals in the wetland system. It was found that the spatial variations of hydraulic conductivity were large, ranging from 2.97×10^{-5} to 3.54×10^{-3} cm/s in the wetland aquifer (Table 5.1), which indicates the complex structure of the wetland aquifer that would affect the water flow and chemical migration in the study area.

The water table in the upland was about 50 m above the sea level, which is higher than the elevation of the headwater wetland surface. The wetland became the discharge area in the study area. The spatial distribution of average hydraulic potential along the cross section A–B is shown in Fig. 5.2. Based on the hydraulic potential measured in the piezometers, the groundwater flowed horizontally in the upper part (W1 to W2) turned upward in the area between W3 and W7 where the water table was high enough to make a seepage zone. In fact, the largest upward hydraulic gradient of 0.8 m/m was found at W7, indicating the flow path of groundwater in the wetland. It was found that there was little change for temporal distribution of the hydraulic potential pattern, indicating the groundwater leaves the wetland to discharge to surface water downstream.

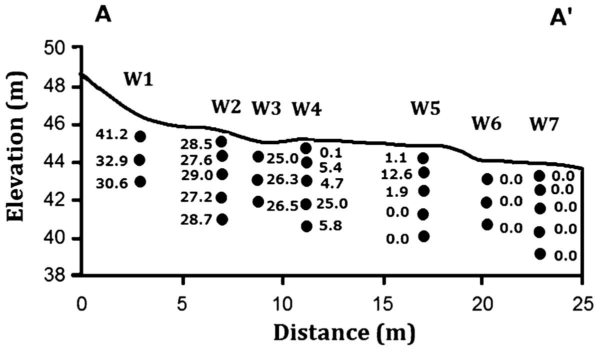

Fig. 5.3 Distribution of average concentration for nitrate (mg/l) in groundwater along A–A′ (based on the data from 2004 to 2007)

5.4.2 Migration of Nitrogen in the Groundwater as Nitrate

Several recent studies have found that larger groundwater nitrate reservoirs exist in catchments than previously suspected (Williams et al. 1997; Burns and Nguyen 2002). Because there were no agricultural activities in the experimental wetland, the nitrate in the aquifers of the study area was expected to derive from the fertilizers used in the upland from which it was transported to the wetland by groundwater. As a result, the groundwater entering into the wetland had the same hydrochemical characteristics as the groundwater in the upland, with a high nitrate concentration ranging from 30.6 to 41.2 mg/l (Fig. 5.3; W1).

As already described, the groundwater flowed across the surface as a seepage or a piping flow in the area between W3 and W7 when it flowed through the wetland. Figure 5.3 shows the distribution of average NO_3^- concentrations along the cross section A–A′ at the study area. Before groundwater entered the wetland, NO_3^- concentrations in the groundwater were high (Tang et al. 2004). When it moved across the wetland, there were three stages for NO_3^- in the groundwater. The first was the high nitrate stage in the area from W1 to W3 where the concentration of NO_3^- was high; the second was the declining stage from W3 to W5 where the concentration of NO_3^- decreased sharply; and the third was a nitrate-free stage from W6 to W7 where NO_3^- was not detectable (Fig. 5.3). Actually, the first stage was located at a zone where the groundwater flowed horizontally, whereas the second and the third stages were found in the zone with the upward groundwater flow. From W3 to W5, the concentration of nitrate in the groundwater decreased dramatically to one half at 1 m in depth or even was not detectable from the piezometers at 3 m below the surface. The decline of nitrate concentration in groundwater was also found in the piezometers at 0.5 m in depth from W2 to W5, which indicated that nitrate removal had occurred in a very narrow zone before groundwater moved up to the seepage zone (Fig. 5.3). Therefore, nitrogen removal in the subsurface may be directly influenced by the structure of aquifers, wetland hydrology (e.g., soil saturation, groundwater flow paths), and subsurface biogeochemistry (organic carbon supply, high nitrate inputs) through cumulative effects on microbial denitrification activity.

Table 5.2 N_2O flux, concentrations of dissolved N_2O, and nitrate in the experimental area (July 6, 2007)

Site	N_2O flux (μg-N/m^2/day)	Dissolved N_2O-N (μg/l)	NO_3-N (mg/l)
W3	1.54	3.61	7.56
W4	0.99	1.21	1.86
W5	0.04	9.38	0.70
W6	0.06	1.25	Undetectable

5.4.3 Migration of Nitrogen in the Groundwater as Nitrous Oxide

Nitrogen, predominantly as nitrate, moves through and leaves wetland systems. NO_3^- can be denitrified in the aquifer, and N_2O is produced as an intermediate product that dissolved in the groundwater. Then, the dissolved N_2O that moves with the groundwater can reach the atmosphere either by upward diffusion to the unsaturated zone or by convective flow to seepage, springs, streams, and wells. In practice, many investigations for N_2O emissions in the terrestrial system are based on the results of N_2O flux measured on the land surface, whether it was N_2O emissions arising at the wetland soils, or that from groundwater and surface waters flowing through the wetland. The knowledge of these emissions is limited because investigations of N_2O behaviors in aquifers are rarely undertaken.

Table 5.2 shows N_2O flux from the surface as well as the concentrations of dissolved N_2O and nitrate in the wetland groundwater. It was found that N_2O flux in the headwater wetland was from 0.04 to 1.54 mg/m^2/day, which was much lower than that in arable lands of the upland (Goto 2012). Also, the concentration of dissolved N_2O in groundwater ranged from 1.21 to 9.38 μg N/l, which was much higher than the value of 0.3 μg N/l in the ambient atmosphere. When groundwater flowed in the wetland aquifer toward the surface, the water pressure decreased continually. This process would help the release of the dissolved N_2O from seepage water on the wetland surface. It was found that the denitrification ability for the wetland in winter season was only about 18.5 % of that in summer in term of N_2O because the temperature of groundwater was one of the important factors controlling the denitrification process in the wetland (Kanno 2008). The dissolved N_2O decreased to 1.25 μg N/l at W6 where nitrate became undetectable. The occurrence of denitrifying microorganisms under wet conditions enabled denitrification processes and might lead to a reduction of N_2O before it could release into the ambient atmosphere. In fact, the dissolved N_2O was three orders less than that of nitrate in the groundwater, indicating the denitrifying microorganisms were active enough to enable denitrification processes and reduce nitrite to N_2 through a reduction of N_2O. As described in Sect. 5.4.1, the water table was too shallow to have an unsaturated zone in the headwater wetland. The N_2O released from the study area was believed to be the dissolved N_2O transported by groundwater rather than that produced at the topsoil between W4 to W7, where no more denitrification should be

Table 5.3 Vertical averages of anion concentrations and hydrochemical parameters at sites W1 to W7 (based on data from 2004 to 2007)

Site	pH	EC (μS/cm)	ORP (mV)	DO (mg/l)	NO_3^- (mg/l)	Cl^- (mg/l)	SO_4^{-2} (mg/l)	HCO_3^- (mg/l)	N (%)	C/N
W1	6.72	135.27	286	6.69	23.67	12.24	1.21	61.63	0.10	11.2
W2	6.76	117.88	265	5.61	31.98	14.68	3.60	48.88	0.10	8.3
W3	6.79	124.37	217	4.58	26.09	13.94	5.33	50.50	0.30	7.5
W4	6.74	100.51	175	3.57	5.25	13.03	10.66	44.90	0.34	7.0
W5	6.57	105.84	103	2.22	2.19	13.89	30.70	41.02	0.50	6.3
W6	6.85	97.28	37	1.85	ud	14.52	12.40	43.20	–	–
W7	6.51	96.02	50	2.00	ud	15.81	6.03	45.81	0.55	7.9

ud undetectable, – no data

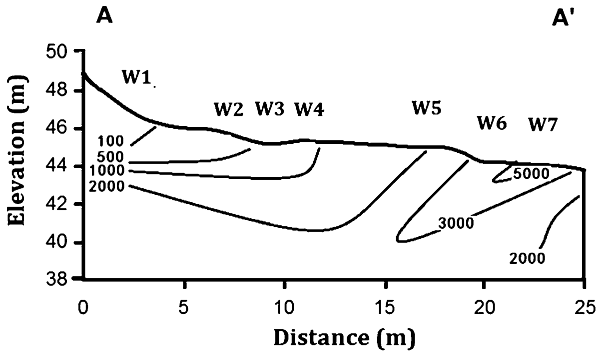

Fig. 5.4 Contour map of number of denitrifying bacteria in groundwater along A–A′ (cells/l)

expected because of the undetectable nitrate (Fig. 5.3). Therefore, it is reasonable that denitrification is the predominant N_2O-producing process and groundwater is the main N_2O carrier in the headwater wetland.

5.4.4 Spatial Distributions of Denitrifying Bacteria in the Experimental Area

The average pH value in the groundwater of the wetland changed from 6.51 to 6.79, which is suitable for the growth of denitrifying bacteria (Table 5.3). The distribution of denitrifying bacteria was controlled by groundwater flow, the redox environment, and nitrate concentration. Associated with nitrate stages described in Sect. 5.4.2, the performance of denitrifying bacteria at each stage indicated their contributions to denitrification processes in the study area. In general, the number of denitrifying bacteria increased from W1 to W7 in the range of about 1×10^2 cells/l to more than 5×10^3 cells/l, which was strongly consistent with nitrate removal in the wetland (Fig. 5.4).

An increase of denitrifying bacteria was found in the area from W1 to W3 at the first stage, where the denitrifying bacteria accounted for 1×10^2 to 2×10^3 cells/l and the nitrate concentration was up to 30 mg/l. More than 5×10^2 cells/l of denitrifying bacteria were seen in the area from W3 to W5, with a sharp decrease of NO_3^- at the second stage. It was found that the number of denitrifying bacteria increased with depth, associated with the decline of nitrate in the groundwater. Especially, up to 1×10^2 cells/l were counted in the aquifer 3 m below the surface at W5, where the nitrate was undetectable. At the third stage, the number of denitrifying bacteria was up to the maximum value without detectable nitrate from the groundwater as shallow as 0.5 m in depth at W7. Furthermore, it was found that the groundwater had more denitrifying bacteria than did the topsoil. It should be considered the growth condition for denitrifying bacteria in the aquifer was much better than in the topsoil.

5.5 Discussion

5.5.1 Characteristics of the Physicochemical Environment in the Wetland

For convenience of explanations, the vertical average concentrations of DO, Cl^-, SO_4^{2-}, and NO_3^- in the study period from 2004 to 2007 were calculated. Table 5.3 shows the variations of average concentrations of DO and anions from W1 to W7. The concentration of NO_3^- decreased from W1 to W5 and was undetectable at the area from W6 to W7. SO_4^{2-} concentration was as low as 1.21 mg/l at W1 when the groundwater flowed into the wetland, increased to the peak value of 30.7 mg/l at W5, and decreased to 6.03 mg/l at W7. Cl^- concentration changed from 12.24 to 15.81 mg/l, much less than that of NO_3^- and SO_4^{2-} along the flow path of the wetland, suggesting there should be some chemical or microbiological reactions associated with the variations of nitrate and sulfate in the wetland because Cl^- is often considered geochemically as the ideal tracer for its conservation in the hydrosphere. On the other hand, the average values of ORP were 286 mV at W1, 103 mV at W5, and 50 mV at W7, respectively. At the same time, DO value decreased sharply from 6.69 mg/l at W1 to 3.57 mg/l at W4 and maintained its value around 2.0 mg/l in the area from W5 to W7 when the groundwater flowed through the wetland (Table 5.3).

DO is thermodynamically preferred as a terminal electron acceptor. When DO depletes, nitrate will replace it as electron acceptor, then denitrification will begin under the low ORP condition. In the denitrification processes, the final products are nitrogen gas (N_2) and a small fraction of nitrous oxide (N_2O) as an intermediate (Kolle et al. 1983):

$$5C + 2H_2O + 4NO_3^- \rightarrow 2N_2 + 4HCO_3^- + CO_2 \tag{5.1}$$

The increasing concentration of SO_4^{2-} with decrease of DO in the shallow groundwater suggests that sulfur- and iron-containing minerals are participants in the redox reactions as

$$4FeS_2 + 15O_2 + 14H_2O \rightarrow 4Fe(OH)_3 + 16H^+ + 8SO_4^{2-} \tag{5.2}$$

The groundwater flow paths are vital channels providing nitrate for denitrifying the bacteria community in the wetland. When groundwater with a high concentration of nitrate flows into the wetland, the microbial denitrifier community extends with DO depletion. Along the pathway, nitrate is transferred by the upward groundwater flow to survive the microbial community in the aquifer. In spite of a small increase of DO at the seepage zone, groundwater brings the denitrifying bacteria from aquifer to the surface, causing the increase of denitrifying bacteria number at the topsoil even if little nitrate is detectable (Table 5.3, Fig. 5.4). From W1 to W5, the negative linear relationship between nitrate and sulfate should be explained by denitrification because of the oxidation for sulfides:

$$14NO_3^- + 5FeS_2 + 4H^+ \rightarrow 7N_2 + 10SO_4^{2-} + 5Fe^{2+} + 2H_2O \tag{5.3}$$

Chemoautotrophic denitrification causes the decline of NO_3^- and the increase of SO_4^{2-} in the water at the same time. When nitrate depletes at the area downstream of W5, sulfate and iron replace nitrate as the new electron accepters in the groundwater of the wetland system. SO_4^{2-} reduction in shallow groundwater is most likely in response to anaerobic respiration accepting energy in a lower DO condition from organic carbon that serves as the electron donor for the microbial desulfurication:

$$SO_4^{2-} + 2C + 2\,H_2O \rightarrow 2HCO_3^- + H_2S \tag{5.4}$$

In general, the carbon used for denitrification by denitrifying bacteria comes from the groundwater and aquifers. Table 5.3 also gives the variation of organic carbon, total nitrogen, and C/N ratio in the wetland. The C/N ratio ranges from 6.3 to 11.2 with an average of 8.1, which is suitable for the growth of denitrifying bacteria along the cross section A–A′ in the study area. At the same time, the low N content in the aquifer promotes the denitrifying bacteria to transform nitrate to nitrogen gas under the oligotrophic condition. As a whole, groundwater provides the suitable condition for the microbial community to survive in the aquifer and develops the reduction sequence of DO, nitrate, and sulfate along the pathway in the wetland.

5.5.2 Effects of Ecohydrological Function on Nitrogen in the Groundwater of the Wetland

Hydrogeomorphologically, the development of the headwater is controlled by the groundwater and geology. The top layer in the wetland should be a mixture of materials transported from the valley slope. This process not only builds a channel to let groundwater transport chemicals but also forms the carbon pool where bacteria can grow. Basically, the farmlands contribute to the nitrate contamination in groundwater by leaching of fertilizers with the infiltration of rainfall in the upland.

The slope provides ample C and N sources that have accumulated over the long term including trees and the underlying litter on the forest floor. The active C and N that is stored on the surface of the slope finally moves downward to the adjacent headwater wetland where the nutrition pool can become available for the growth of bacteria.

When the groundwater flows through the headwater wetland, the redox potentials decrease with the removal of dissolved oxygen in the aquifers as described in Sect. 5.5.1. As a result, nitrate functions as an electron acceptor and is reduced to N_2 in the process of denitrification. At the same time, transformations between species of nitrogen are almost exclusively facilitated by microorganisms. Thus, the equilibrium relationships must be controlled by microbial kinetic factors. The dissolved N_2O can accumulate in the groundwater when gaseous diffusion is restricted in the wetland. When groundwater flowed from W3 to W6, NO_3-N decreased from 7.56 mg/l at W3 to undetectable at W6, and N_2O-N changed from 3.61 μg/l at W3 to 1.25 μg/l at W6, with the highest value of 9.38 μg/l at W5 (Table 5.2). Based on the data in the study area, denitrification is identified mainly in the area from W3 to W7. Because the concentration of nitrate is much higher than that of the dissolved N_2O, nitrate becomes the dominant species of nitrogen in the groundwater of the study area. It is reasonable to consider that the total inorganic nitrogen could be estimated as the sum of N_2O-N and NO_3-N in the groundwater. The nitrate entering the wetland from the recharge area eventually leaves the wetland as nitrogen gas or N_2O are brought to the seepage zone by the groundwater because of denitrification in the aquifers.

Water is one of the key factors in the ecosystem, and awareness of ecosystem services depends on the type of services, the user, and the spatial scale of its delivery. For example, fertilizers ensure good crop yields and return in the upland, but also lead to nitrogen contamination of surface and groundwater and eutrophication of aquatic ecosystems in the lowlands downstream. Fortunately, the headwater wetland can be considered as a transformer between the biogeochemical source and sink for nitrate, which maintains the ecosystem services available to humans at both upland and lowland because of the ability for denitrification. As a carrier for chemicals in the watershed, groundwater in the study area is recharged at the upland, then moves down the valley side and flows up to the surface as spring or seepage at the headwater wetland. Therefore, the role of wetland as a source, sink, or transformer of materials depends largely on hydrodynamics and geochemistry.

Also, there is often a spatial and temporal mismatch between the places where humans use ecosystem services, the location of ecosystems that produce them, and the origins of stressors and drivers that modify the services. The residence time in the watershed is the key to understanding hydrodynamics that is especially important in the exchange of materials between the wetland and upland ecosystems. The age of seepage and spring in the headwater wetland was more than 20 years as estimated by the dating technique with SF_6 (Hashimoto et al. 2009). Therefore, the nitrate in the wetland groundwater should be at least that leaching from the upland farmlands in past decades rather than at present.

5.6 Conclusions

Hydrogeomorphologically, the headwater wetlands exist at the interface between upland and lowland environments. Wetland hydrology was the driving force that determined soil development, the assemblage of plants and animals that inhabited the site, and the type and intensity of biochemical processes. The various microbiological, chemical, and physical properties of the aquifer that influenced N_2O emissions and nitrate removal were distributed throughout the headwater wetland. NO_3^- concentration decreased in the area from W1 to W5, and was not detectable at the area downstream of W6, when the groundwater flowed through the wetland. Associatively, DO and ORP values decreased sharply, ranging from 1.85 to 2.22 mg/l and from 37 to 103 mV in the area from W5 to W7, respectively. Depending on the reduction environment prevailing in the wetland aquifer, the number of denitrifying bacteria increased from about 1×10^2 cells/l at W1 to more than 5×10^3 cells/l at W7, which was firmly consistent with nitrate removal along the wetland. It was found that N_2O formed as an intermediate product in the denitrification processes and then dissolved in the groundwater. The N_2O released from the wetland was the gas dissolved in the groundwater rather than that produced at the wetland topsoil. Because the concentration of nitrate was much higher than that of the dissolved N_2O, NO_3-N was the dominant species of nitrogen in the groundwater and eventually left as nitrogen gas or N_2O from the headwater wetland. Considering the ecohydrological functions of the wetland, groundwater also provides a dynamic system for the delivery of chemicals through the aquifers, and the headwater wetland plays an important role in the removal of nitrate coming from the upland and in improvement of water quality for the river downstream.

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Chapter 6
Numerical Simulation of Groundwater Flow Using Stable Isotopes of Oxygen and Hydrogen as Natural Tracers

Shinji Nakaya

Abstract Stable isotopes of oxygen and hydrogen have the potential to serve as tracers for both source and flow paths in a groundwater system. The ratios of stable isotopes of oxygen ($\delta^{18}O$) and hydrogen (δD) can be used as natural tracer parameters to separate multiflow groundwater paths by applying a simple inversion analysis method to determine the differences between observed and calculated $\delta^{18}O$ and δD data in a simple mixing model. The model presented here assumes that the distribution of natural tracers in the steady state is governed by simple mixing between flow paths with a normal distribution of flow rate. When the inversion analysis and simple mixing model were applied to the multiflow system of the Matsumoto Basin, which is surrounded by Japanese alpine ranges, the end-members of the relationship between observed $\delta^{18}O$ and δD could be separated spatially into specific groundwater flow paths in the multiflow system of shallow and deep groundwater flow paths.

Keywords Groundwater flow path • Inversion analysis • Numerical simulation • Simple mixing • Stable isotopes of oxygen and hydrogen

6.1 Introduction

In recent global issues concerning the lack of water and water pollution, people have fully recognized that the value of groundwater becomes higher for human life and that a sustainable use of groundwater is desired. However, people do not understand subsurface groundwater flow in basin scale because it is invisible. Therefore, we need to understand groundwater flow in the water cycle and to make groundwater flow visible as is a river network. Understanding groundwater

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recharge patterns and basin-scale flow paths is important for the management of water resources and environmental conservation in urban areas. In a basin-scale area, especially one surrounded by mountain ranges, a wide range of recharge waters arise from potential recharge areas such as mountains and plateaus through different aquifers. Even for governments that must manage the water resources for people, it will be a tough task to delineate a groundwater flow system through transboundary aquifers. More simple and reliable methods for giving basic data to trace and to manage water resources are expected.

The principal chemical constituents of groundwater can be used to characterize the water quality of different aquifers. Principal component analyses (PCA) have been employed to trace the origin of the groundwater and calculate the mixing portions and mass balances from ambiguous groundwater composition data (e.g., Laaksoharju et al. 1999). These PCAs make it possible to evaluate the spatial and temporal variations in the principal physicochemical processes implicated in groundwater quality (e.g., Sanchez-Martos et al. 2001).

Previous researchers have investigated the stable isotopic ratios of oxygen ($\delta^{18}O$) and hydrogen (δD) to study groundwater recharge and flow because these stable isotopes of oxygen and hydrogen have the potential to trace both source and flow paths in groundwater systems (Craig 1961; Mizutani and Oda 1983; Mizota and Kusakabe 1994; Mizutani and Satake 1997; Adams et al. 2001; Mizutani et al. 2001; Weyhenmeyer et al. 2002; Lui et al. 2004; Wilcox et al. 2004; Komiya et al. 2003; Nakaya et al. 2007). However, few studies have addressed the potential contribution of $\delta^{18}O$ and δD data to the separation of groundwater flow paths from a multiflow system at the basin scale. Most previous research, in which hydrologic and physical approaches were employed to measure parameters such as potential head, flow velocity, and geophysical techniques, have been found to be ill suited to the estimation of the spatial distribution of $\delta^{18}O$ and δD data to identify groundwater flow paths. The spatial estimation method, however, does not provide for the objective diagnosis of spatial separation of flow paths. It will be fruitless to separate the flow paths spatially using hydrologic and physical approaches without chemical data in the groundwater mixing system (Nakaya et al. 2007). However, determining specific groundwater flow paths in groundwater systems allows the investigation of hydrologically important problems such as water resources management, environmental conservation, and the mechanism of pollution.

Groundwater flow path evaluations often use a conservative species in water as a tracer. A stable isotope is promising natural tracers in groundwater, because of their conservative behavior in an aquifer. The study of groundwater mixing with two tracer parameters (T1, T2), such as concentrations or stable isotopic ratios, is indispensable for separating flow paths in multiflow systems within basins. Carrera et al. (2004) have proposed a methodology for computing mixing ratios with uncertain end-members by maximizing the likelihood of concentration measurements with respect to both mixing ratios and end-member concentrations. Chemical mass-balance mixing methods identify and mix two or more end-member waters with distinct chemical compositions to determine the chemical compositions of sample waters (Maule and Stein 1990; Laaksoharju et al. 1999; Pitkanen et al. 1999; Joerin et al. 2002; Carrera et al. 2004; Laudon et al. 2004; Lui et al. 2004).

In "Numerical simulation of groundwater flow" of the lectures held as UNESCO's International Hydrological Programme (IHP) "Groundwater as a key for adaptation to changing climate and society," the twentieth IHP Training Course 7–20 November, 2010 in Nagoya and Kyoto, Japan, a simple model was introduced by Nakaya to examine whether the calculation of $\delta^{18}O$ and δD can explain observed data, based on the simple mixing of flow paths (Nakaya et al. 2007). This chapter, based on Nakaya et al. (2007), reflects the part of the simple simulation method in the lecture. First, a simple method is presented to spatially separate the specific groundwater flow paths from a multiflow groundwater system using $\delta^{18}O$ and δD as natural tracer parameters within a simple mixing model. The method is then applied to the observed $\delta^{18}O$ and δD groundwater field data in a basin for the purpose of demonstrating the validity of the proposed method by spatially separating the specific groundwater flow paths.

6.2 Model

Mixing of groundwaters in a multi-groundwater system appears as plots for sampling point on the δ-diagram. Figure 6.1a schematically shows a case of two groundwaters mixing. Groundwaters A and B flow along the flow paths A and B, respectively, and mix with each other. When both end-members of flow path are A and B with each ($\delta^{18}O$, δD) set, the result of two groundwaters mixing appears as

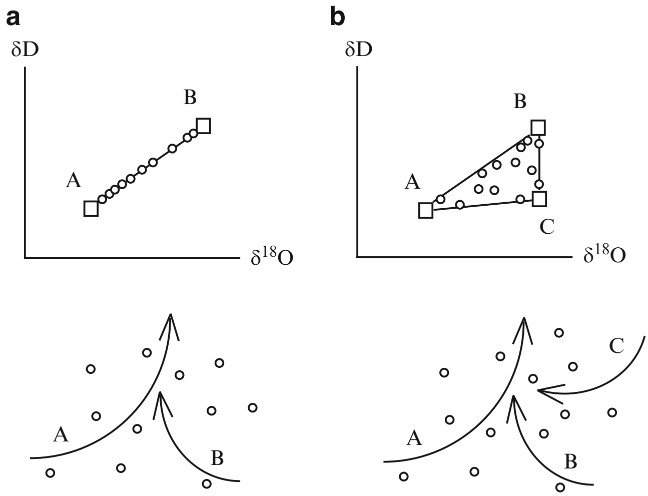

Fig. 6.1 Schematic diagram of groundwater mixing on δ-diagram and spatial distribution of sampling points in a two groundwaters mixing case (**a**) and a three groundwaters mixing case (**b**)

plots on a line, which indicates the ($\delta^{18}O$, δD) value of each sampling point, with end-members A and B on δ-diagram incompletely mixing. Similarly, in a case of three groundwaters mixing, the results are plotted within a triangle with three end-members A, B, and C (Fig. 6.1b).

Mixed groundwater with *n* end-member sources will generally plot within an *n*-sided polygon on a $\delta^{18}O$ versus δD graph (δ-diagram) (Fig. 6.1). An end-member for ($\delta^{18}O$, δD) represents a groundwater source and has the potential for fundamentally tracing a groundwater flow path. Recharge water giving an end-member is likely to flow along a large-scale flow path from a horizontal or vertical outer boundary, which causes channeling of the tracer in heterogeneous geologic media, with some spread (Anderson 1997). Several intensive inflows corresponding to end-members can be assumed on the boundary. In general, local groundwater at a sample point is formed as a result of multiflow mixing with neighboring flow paths. Therefore, if the process of groundwater mixing with multiflow paths is solved using $\delta^{18}O$ and δD as conservative parameters, the flow paths will be separated by the inversion method. The fundamental view of the inversion method is that a concentration of a conservative parameter exists at a local point as a result of simple multiflow mixing with neighboring flow paths. The model concentrations are calculated at sample points and then the location of each flow path and the specific parameters are determined to adequately fit the calculated concentrations to the observed concentrations.

The estimation of local flow direction provides useful data for the modeling of groundwater flow paths. When a tracer moves through a porous medium along a flow path, the direction of the minimum gradient of the tracer concentration is likely to be identical to the direction of local maximum flow rate, because the tracer advances in the direction of the flow even in a uniform flow field (Bear 1972). In the present study, the direction of the minimum gradient of the stable isotopic ratio δ ($\delta^{18}O$, δD) is assumed to be spatially identical to the local flow direction in a multi-groundwater flow system because the stable isotopes, O^{18} and H^2, are expected to behave as natural tracers. The sample points grouped around an end-member *j* on an observed $\delta^{18}O$ versus δD graph are likely to be spatially distributed around or along the flow path *j*.

6.2.1 Numerical Modeling of Simple Mixing

The heterogeneity of geologic media is based on fractures, macropores, and other geologic structures. These pore structures, which form heterogeneous hydraulic conductivity fields, affect fluid flow and solute transport at the macroscopic level by creating nonuniform flow velocity fields (Dagan 1982; Gelhar and Axness 1983). The random character of the velocity distribution for a homogeneous fluid in a spatially averaged uniform flow through a homogeneous medium is a result of the nature of the porous matrix (Bear 1972).

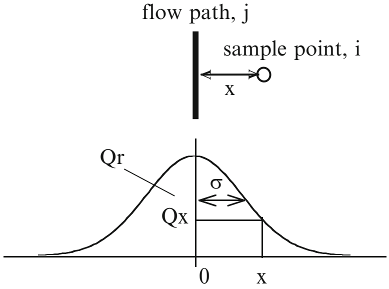

Fig. 6.2 Normal distribution of flow rate Qx for a flow path j. (From Nakaya et al. 2007)

Anderson (1997) reviewed investigations of fluid flow and solute transport in heterogeneous geologic media and macroscopic heterogeneities in the subsurface. Such phenomena are often referred to as macroscopic preferential flows (Moreno et al. 1988; Ewing and Gupta 1993). Field and laboratory observations show that water and chemicals often move through media along preferred pathways (Bourke et al. 1985; Pyrak et al. 1985; Adams and Gelhar 1992). Preferential flow can result in the rapid movement of fluids and chemicals from upstream to downstream regions or from the ground surface to the water table. A flow model that describes flow in both heterogeneous as well as in homogeneous geologic media should therefore be considered.

In the present study, each groundwater flow giving an end-member with the same sources is hypothesized to be a flow path with a normal flow rate distribution in the direction (x), perpendicular to the flow path (Fig. 6.2):

$$Q_x = \frac{Q_r}{\sqrt{2\pi\sigma}} e^{-\frac{x^2}{2\sigma^2}}, \tag{6.1}$$

where Q_x ($=Q(x)$) and Q_r are the flow rate at distance x and the total flow rate in the plane perpendicular to the path, respectively. σ represents the standard deviation of Q_x. This type of flow model can accommodate both preferential flow paths as well as spatially averaged uniform flows using the variable σ, because a smaller σ value can describe a preferential flow path and a larger σ can approximate spatially averaged uniform flow excluding radial flow. Few investigations report whether the σ value is constant or fluctuates largely along a flow path. In this study, σ is hypothesized to be constant for each flow path to avoid complications in calculation.

When n number of groundwaters for path j, with different sources of δ^{18}O and δD, are mixed homogeneously (Fig. 6.3), the ratio of the stable isotope δ_i for oxygen or hydrogen (δ^{18}O or δD) at sample point i can be calculated using Eq. (6.2):

$$\delta_i = \frac{\sum Q_{xij}\delta_j}{\sum Q_{xij}}, \tag{6.2}$$

where Q_{xij} ($=Q_{ij}(x)$), obtained in Eq. (6.1), is the flow rate of path j at sample point i at a distance x from path j, in the direction perpendicular to path j. δ_j is the δ^{18}O or δD of path j.

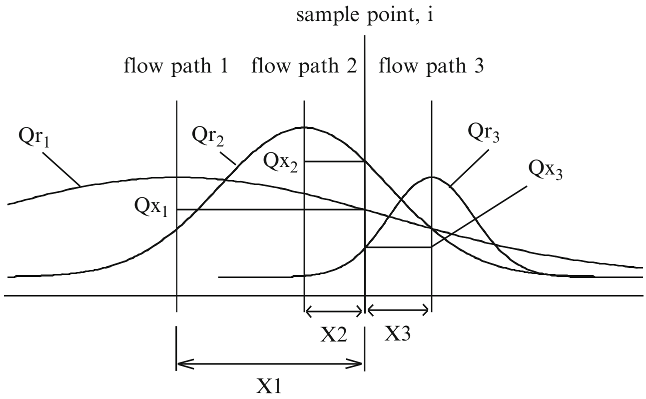

Fig. 6.3 Simple mixing of groundwater flow paths. (From Nakaya et al. 2007)

6.2.2 Inversion of Flow Path

A nonlinear least squares inversion is performed using a modified Marquardt method (Marquardt 1963; Fletcher 1971). The nonlinear least squares solution can be obtained by minimizing the residual sum of squares between the observed δ_i^{obs} and the calculated δ_i^{cal} in Eq. (6.2) at sample point *i*:

$$\sum_{i=1}^{m} \left[\delta_i^{obs} - \delta_i^{cal}\right]^2 \rightarrow \text{minimum}, \tag{6.3}$$

where *m* is the total number of sample points.

Practically, the spatial distribution of path *j*, and δ^{18}O, δD, Q_r, and σ of path *j*, can be obtained as a nonlinear least squares solution with a nonnegative constraint of Q_r and σ, according to procedures (6.1)–(6.4) of the inversion analysis listed below (Fig. 6.4).

1. From the plots of observed δ^{18}O versus δD of groundwater samples, *n* tentative end-members are determined to obtain the number of flow paths, *n* (Fig. 6.5). In the present study, the tentative end-members for each groundwater system are selected based on the fundamental rule that (δ^{18}O, δD) is the end member of a sample point, giving an apex of an *n*-polygon, which contains whole (δ^{18}O, δD) data, on the δ-diagram and the location of the end-member (Fig. 6.6a–d). The sample point giving the tentative end-member lies along a flow path within the area spatially closer to a recharge zone or a source.
2. The flow path *j* (*j* = 1, 2, 3,. . ., *n*) is spatially set as a curved line [with normal distribution of flow rate from Eq. (6.1)] composed of several line elements with nodal points. In most cases where δ^{18}O and δD data are acquired at many sample points, the tentative flow paths can be estimated using the pathway that is most

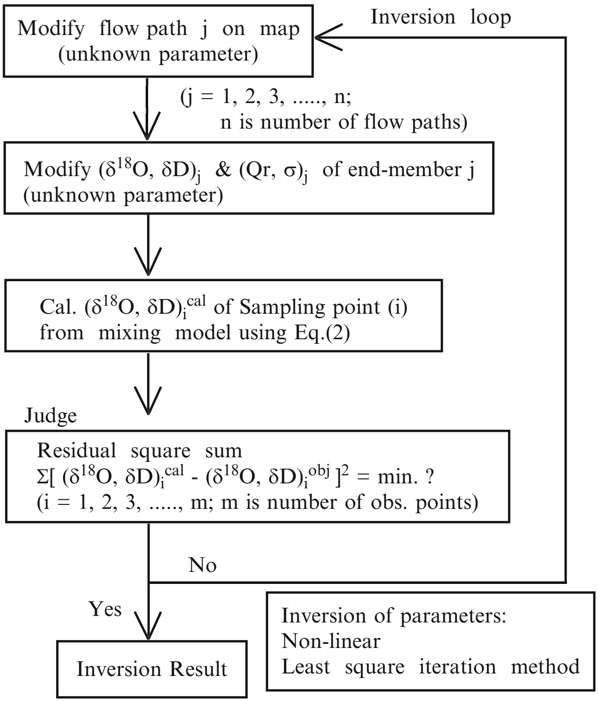

Fig. 6.4 Flowchart of inversion analysis for spatial separation of groundwater flow paths. *obs.*, observed

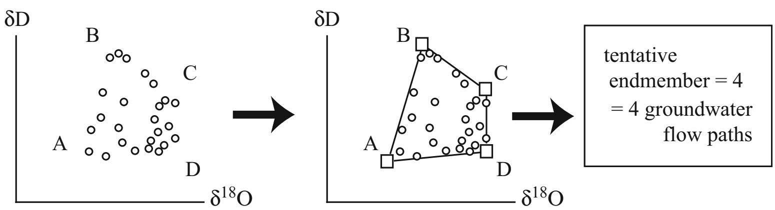

Fig. 6.5 Determination of tentative end-members and number of flow paths from the plots of observed δ^{18}O versus δD of groundwater samples

likely to be parallel to the local flow directions, through the spatial locations of sample points grouped around the tentative end-member j on the observed δ-diagram, referring to the local flow directions estimated at the sample points (Fig. 6.6e, f). From the following equation, the local flow directions give the gradient minimum of δ at each sample point i:

$$\delta^{i}_{\text{gradient}} = \left\{ \left[\delta^{18}\text{O}^{\text{obs}}_{i} - \delta^{18}\text{O}^{\text{obs}}_{k} \right]^2 + \left[\delta\text{D}^{\text{obs}}_{i} - \delta\text{D}^{\text{obs}}_{k} \right]^2 \right\}^{1/2} \Big/ L_{i,k}, \qquad (6.4)$$

$$(k = 1, 2, 3, \ldots, m_i, k \neq i, m_i \geqq 3)\ ,$$

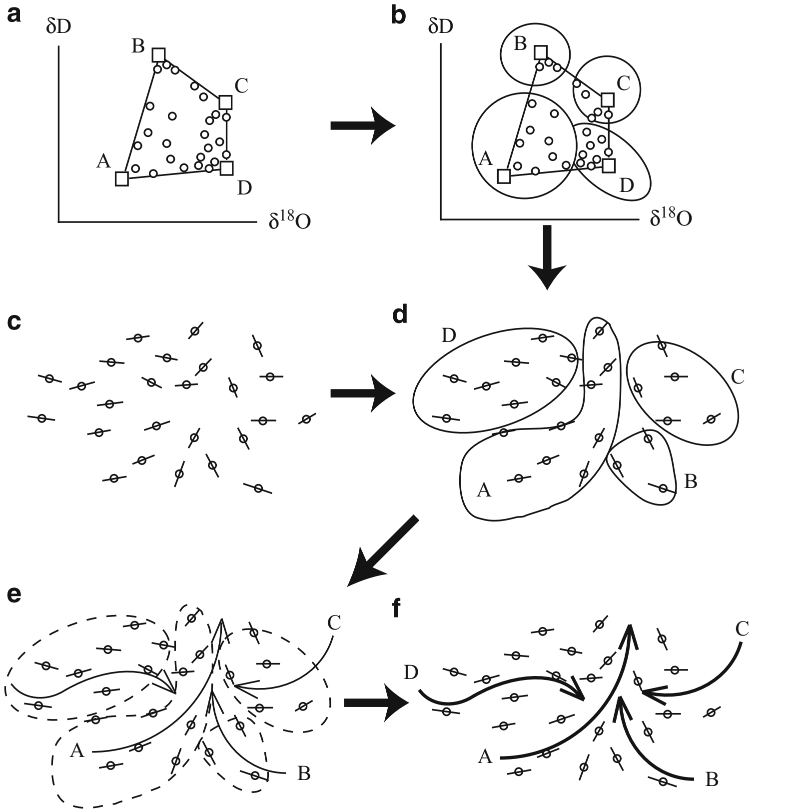

Fig. 6.6 Procedure of constructing of initial model for inversion analysis. (**a**) Determination of tentative end-members on the δ-diagram. (**b**) Grouping of sample data including each tentative end-member on the δ-diagram. (**c**) Local flow directions estimated from the gradient minimum of δ at each sample point. (**d**) Grouping of sample data including each tentative end-member on map. (**e**) Tentative flow paths estimated from local flow directions and spatial locations of sample points grouped around a tentative end-member. (**f**) Tentative flow paths determined from (**a**) to (**e**)

where $\delta^{i}_{\text{gradient}}$ is the gradient combined with gradients of δ^{18}O and δD at sample point i, $\delta^{18}\mathrm{O}_k^{\text{obs}}$ and $\delta \mathrm{D}_k^{\text{obs}}$ are the observed δ^{18}O and δD at sample point k, respectively, $L_{i,k}$ is the distance between sample points i and k, m_i is the number of sample points within a constant distance d from i, and, in this study, d was determined from the maximum distance, which m_i is equal to 3. The tentative flow paths are modified in the inversion analysis.

3. The tentative values for δ^{18}O, δD, Q_r, and σ of path j must be set before the start of the inversion analysis using an iterative inversion solver, or modified Marquardt method. The nodal point coordinates, δ^{18}O, δD, Q_r, and σ of path j, are set, with (δ^{18}O, δD) as a tentative end-member on the δ-diagram, 1.0 km^3/year and 1.0 km are used as initial values for the inversion, and then iteratively modified in the

inversion analysis to solve Eq. (6.3) because Q_r of each flow path is determined as a relative value for the final inversion solution in the case of no data for Q_r. The optimum values for $\delta^{18}O$, δD, Q_r, and σ of path j are obtained against the tentative flow paths, and the performance of the inversion is checked. If the spatial distribution of the tentative flow paths is irrelevant, a large residual sum remains. A flow path with a partially irrelevant path can be detected from both the group on the δ-diagram and the spatial distributions of sample points resulting in a large residual between the observed and calculated δ.

4. To obtain optimum flow paths, the location of the partially irrelevant path is added as a parameter to other parameters in the inversion, $\delta^{18}O$, δD, Q_r, and σ of each path before the next inversion analysis is started. This step is repeated until a minimum value is obtained for the residual sum of squares between the observed δ_i^{obs} and the calculated δ_i^{cal}.

6.3 Model Application

6.3.1 Site Description and Hydrogeological Setting

The study area (550–750 m above sea level) for the application of the simple mixing model is located in the Matsumoto Basin, surrounded by the Japan alpine ranges, which exceed 3,000 m above sea level (Fig. 6.7). The Matsumoto Basin is regarded as an area rich in water resources and is the gateway to the Chubu-Sangaku National Park (the Japan Alps). Aquifers in the Matsumoto Basin consist mostly of several conglomerate layers (CL), supplied from the western mountains, which are composed of Paleozoic-Mesozoic formations. The granite and eastern mountains are composed of Miocene formations and diorite. The sediments in the study area are classified into the Enrei Formation, Nashinoki CL, Kataoka CL, Akagiyama CL, Nakayama peat layer, Hata CL, Moriguchi CL, and the alluvial layer, in ascending order (Sakai 1982). The upper three layers are shallow aquifers forming the present-day fan and river terrace; the lower layers are deep aquifers forming a deep unit of the basin and the hill and mountain zone (Fig. 6.7). The Itoigawa–Shizuoka tectonic line (ISTL) extends in the north–south direction, and several faults derived from the ISTL are inferred under the Matsumoto Basin.

6.3.2 Hydrochemical Data and Experimental Procedures

Groundwater samples were collected from depths of 0–200 m in domestic, agricultural, industrial, and urban wells, and from natural springs in 2000 (99 samples, July–October), 2001 (98 samples, July–October), and 2002 (98 samples,

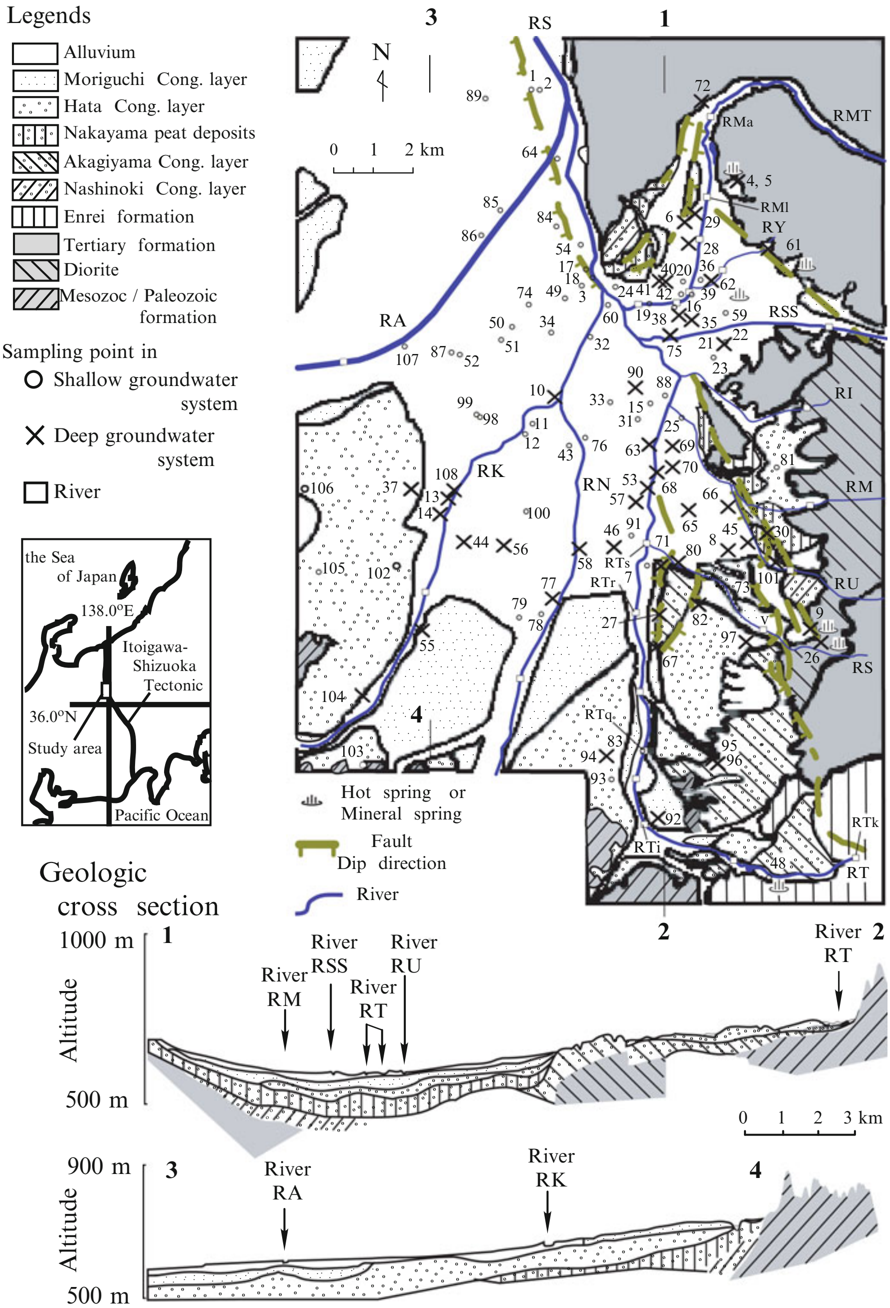

Fig. 6.7 Sampling points and geologic map (modified from Sakai 1982) of study area. (From Nakaya et al. 2007)

July–October). The sample locations, including river water samples, are illustrated in Fig. 6.3. Each sample was measured for the following geochemical species: HCO_3^- at 4.8 alkalinity (in situ), Cl^-, $SO_4{}^{2-}$, and $NO_3{}^-$-N (by ion chromatography), Na^+ and K^+ (by atomic absorption spectrometry), and Ca^{2+} and Mg^{2+} (by titration). The electrical balance was within 3.5 % for each sample. D/H analysis of the samples, mainly those collected in the summer of 2000, was performed using a uranium reduction technique to extract hydrogen from the water (Friedman and Smith 1958), followed by hydrogen isotope mass spectrometry. The $^{18}O/^{16}O$ ratio of the samples, collected mainly in 2000, was determined using the automated CO_2–H_2O equilibration technique (Epstein and Mayeda 1953; Chiba et al. 1985), followed by mass spectrometry (micromass prism). The stable isotope results were expressed in common δ notation, or the proportional deviation (‰) from Vienna Standard Mean Ocean Water (V-SMOW; Coplen et al. 1983). The analytical errors of the δD and $\delta^{18}O$ measurements are ±1.5 ‰ and ±0.1 ‰, respectively.

6.3.3 Initial Groundwater Flow Path Model of Inversion

In this study area, an initial flow path model was set to each shallow and deep aquifer in inversion analysis. For the shallow groundwater system (SGS), the nine initial flow paths were composed of 6 to 12 line-elements with a total of 30 nodal points, and seven initial flow paths were composed of 3 to 8 line-elements with a total of 30 nodal points for the deep groundwater system (DGS). Each flow path independently goes through domains from upstream to downstream. The observation data used in inversion were 51 sets of ($\delta^{18}O$, δD, and coordinates of sampling point locality) measured for the shallow groundwaters and 51 sets of ($\delta^{18}O$, δD, and coordinates of locality) measured for the deep groundwaters.

6.3.4 Results

The shallow and deep groundwaters in the study area have been characterized by Komiya et al. (2003) and Nakaya et al. (2003) according to water quality type, well depth, and geologic structure (Sakai 1982). As shown in Fig. 6.8, the shallow groundwaters (shallower than approximately 50 m below ground surface) in the Matsumoto Basin are almost always of the Ca–HCO_3 type, whereas the deep groundwaters (at depths between 50 m and 200 m below ground surface) are more varied in composition (Ca–HCO_3, Na–(Cl)–HCO_3, Na–SO_4, and Ca–HCO_3–SO_4 types).

The stable isotopes of oxygen and hydrogen have the potential to be applied to the tracing of both flow path and source in groundwater systems. Waseda and Nakai (1983) and Komiya et al. (2003) reported the effects of surface water altitude as

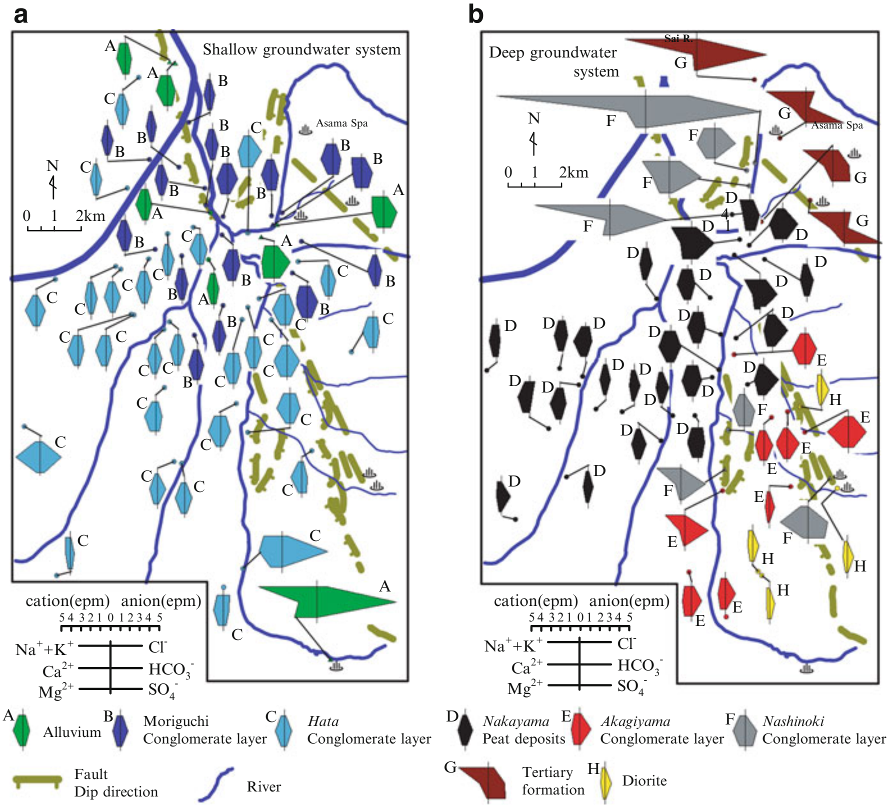

Fig. 6.8 Hexadiagram for shallow groundwaters (**a**) and deep groundwaters (**b**) in study area in summer 2002. (From Nakaya et al. 2007)

-0.25‰ of δ^{18}O per 100 m and in -2.0‰ of δD per 100 m, and the kinetic (latitude–longitude) effect as in 9.1–22.1‰ of *d*-value ($d = \delta D - 8\delta^{18}O$) changes in the meteoric water line (Waseda and Nakai 1983) for central Japan, including the present study area (Fig. 6.9). Figures 6.10a, b and 6.11a, b show the spatial distribution maps of the observed isotope ratios δ^{18}O and δD with the estimated local flow directions providing the minimum gradient for δ at each sample point, $\delta_{gradient}$ from Eq. (6.4) in the shallow and deep groundwater systems, respectively. Figures 6.10c and 6.11c show the groundwater potential contours and the directions of maximum hydraulic gradients at observed points, estimated from 37 and 25 sample data observed in August 1993 (Matsumoto 1993) in the shallow and deep groundwater systems, respectively. The directions of the minimum gradient for δ are concordant with those determined for the maximum hydraulic gradients in both groundwater systems. These results indicate that several groundwater flows assemble to the center of the basin from several recharge zones and then flow to the north.

Figure 6.12 shows the relationship between δ^{18}O and δD (δ-diagram) for the river waters and the shallow and deep groundwater systems. These data are plotted in a

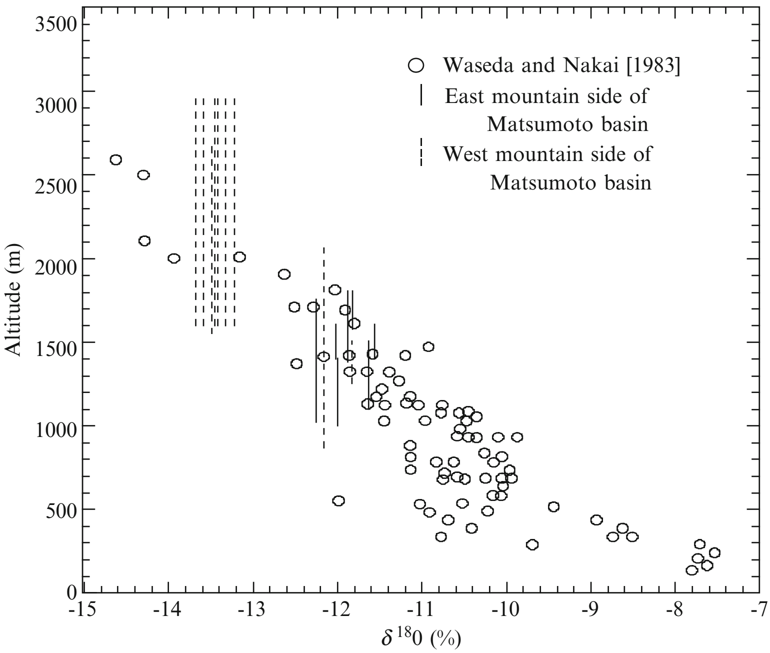

Fig. 6.9 Altitude effects of $\delta^{18}O$ in Central Japan including Matsumoto Basin. (From Waseda and Nakai 1983 and Komiya et al. 2003)

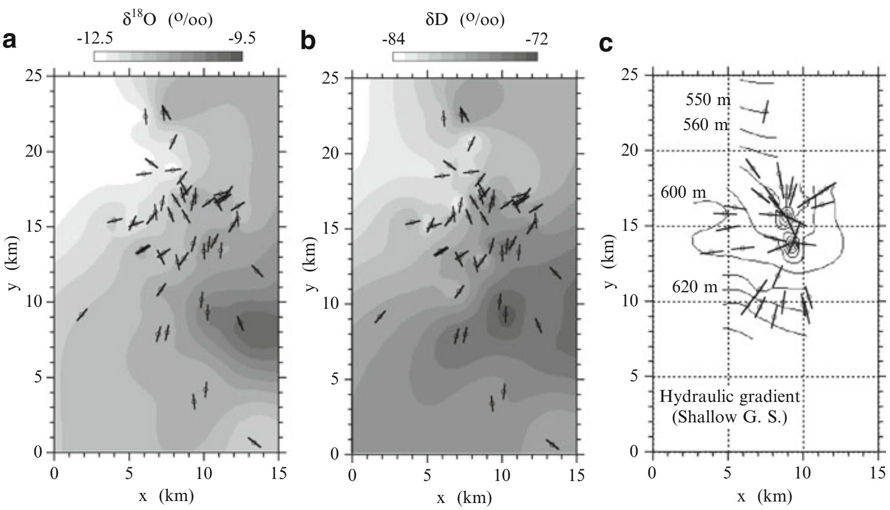

Fig. 6.10 $\delta^{18}O$ (**a**) and δD (**b**) with estimated local directions of minimum δ gradient, and estimated local directions of maximum hydraulic gradient with potential contour lines (**c**) in shallow groundwater system. (From Nakaya et al. 2007)

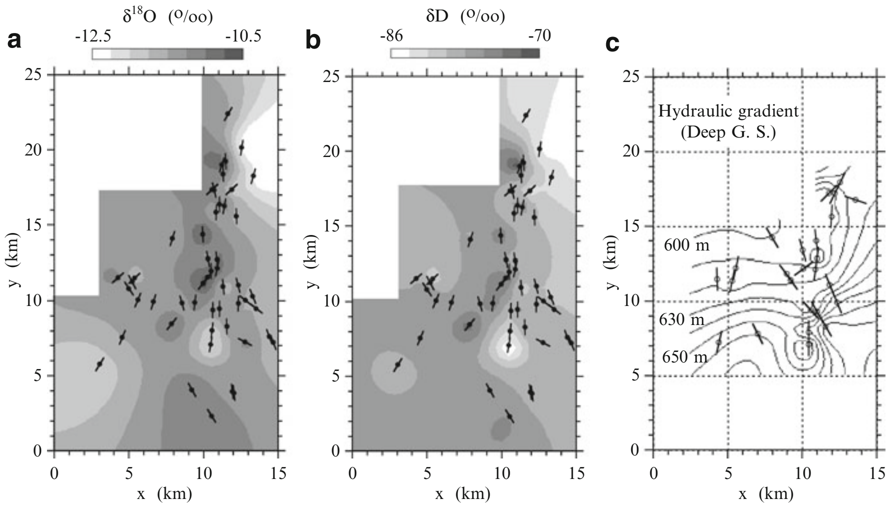

Fig. 6.11 δ^{18}O (**a**) and δD (**b**) with estimated local directions of minimum δ gradient, and estimated local directions of maximum hydraulic gradient with potential contour lines (**c**) in deep groundwater system. (From Nakaya et al. 2007)

widespread area between meteoric water lines ($\delta D = 8\delta^{18}O + 10$) (Craig 1961) and ($\delta D = 8\delta^{18}O + 17$) (Matsubaya et al. 1973). These plots reflect an intricate, kinetic effect between the altitude effect and the effect of distance heading inland from the coasts at the Pacific Ocean or the Sea of Japan (Waseda and Nakai 1983; Mizota and Kusakabe 1994). In the present study, the tentative end-members for each groundwater system are selected based on the fundamental rule that an end-member is the (δ^{18}O, δD) of a sample point, giving an apex of an *n*-polygon, which includes the complete (δ^{18}O, δD) data, on the δ-diagram and the location of the end-member, and the sample point lies closer to a recharge zone or a source (Fig. 6.13). The ten (A, B, C, D, E, F, G, H, J, and K) and nine (A, B, C, D, E, F, G, M, and N) tentative end-members are distinguishable by the differences in δ^{18}O and/or δD, considering the analytical errors and the spatial locations, as each apex of the polygon on the δ-diagram for the shallow and deep groundwater systems in the study area, respectively.

In Fig. 6.13, δ^{18}O and δD are separated into several groups on the δ-diagram, according to the combination of (δ^{18}O, δD) and the sample location. Each member in a group is chosen from the sample points, which are closer to the end-member on the δ-diagram, and the spatial location. The members that lie between the end-members and are close to one another on the δ-diagram and in spatial location are distinct groups such as groups I in both the SGS and the DGS. Figure 6.13c, d show the spatial distributions of sample points grouped around and among the tentative end-members on the δ-diagram (Fig. 6.13a, b) with directions in the minimum of the $\delta_{gradient}$ at each sample point.

According to the spatial distribution of the flow directions estimated from the minimum of the $\delta_{gradient}$ and the groups of the combination of (δ^{18}O, δD), the flow

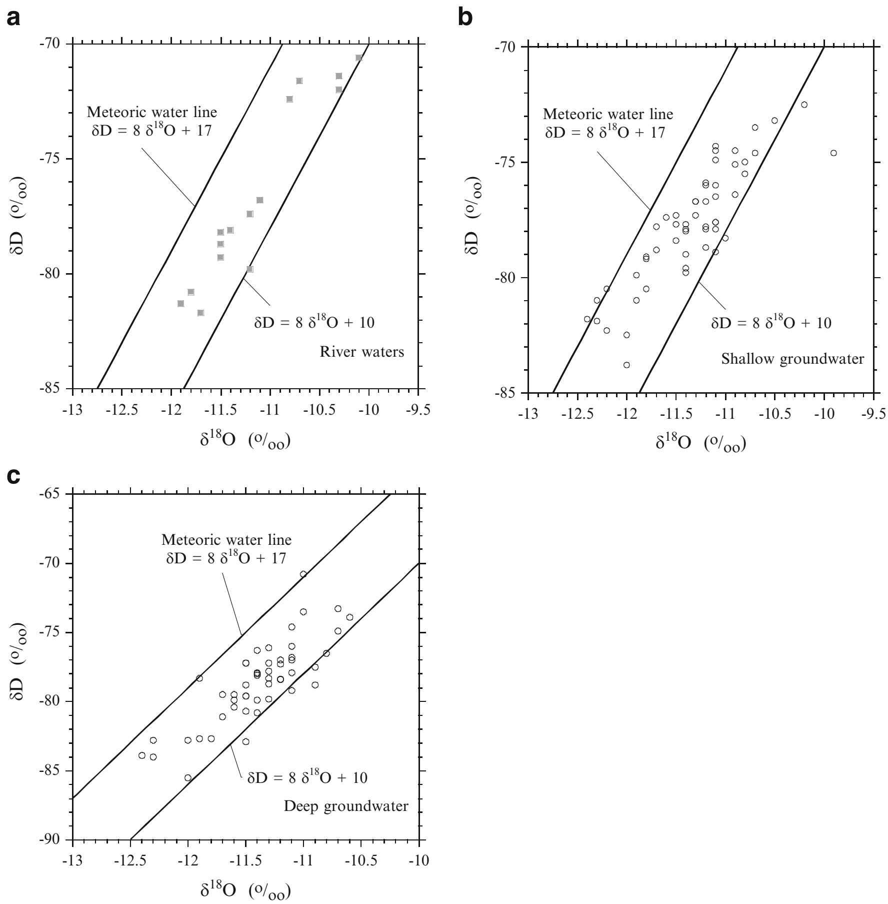

Fig. 6.12 Relationship between δ^{18}O and δD with meteoric water lines (Craig 1961; Matsubaya et al. 1973) for river water (**a**) and shallow (**b**) and deep (**c**) groundwater systems. (From Nakaya et al. 2007)

paths can be estimated more clearly than shown on the contour map of δ^{18}O or δD (Figs. 6.10, 6.11). The tentative flow paths, which are parallel to the local flow directions estimated at the sample points, were estimated from the pathway through the spatial locations of sample points grouped around the tentative end-members on the observed δ-diagram (Fig. 6.12). It was noted that the tentative end-member C (no. 73) in the SGS is isolated on the δ-diagram (Fig. 6.13a) and was omitted from the simple mixing analysis. Therefore, the number of flow paths for the SGS and DGS are set to nine and nine, respectively.

At the beginning of the inversion analyses, the initial flow paths are spatially set from the grouped spatial field distributions of sample points and the estimated local flow directions at the minimum of the $\delta_{gradient}$. The initial values of δ^{18}O and δD for

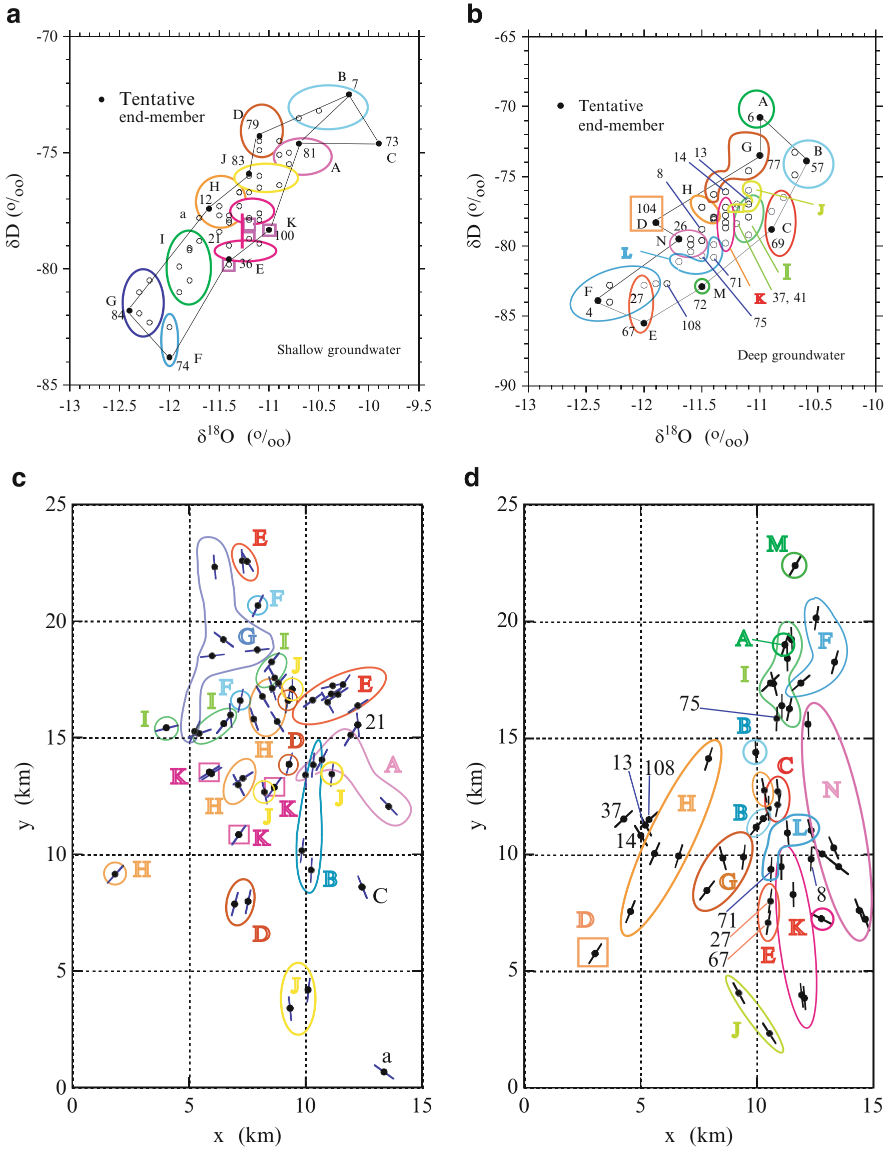

Fig. 6.13 Relationship between δ^{18}O and δD for the shallow (**a**) and deep (**b**) groundwater systems, where the isotopic compositions are grouped. Spatial distributions of sample points grouped around and among the tentative end-members on the δ^{18}O versus δD graph (Fig. 6.8a, b) with directions in minimum of the $\delta_{gradient}$ at each sample point for the shallow (**c**) and deep (**d**) groundwater systems. (From Nakaya et al. 2007)

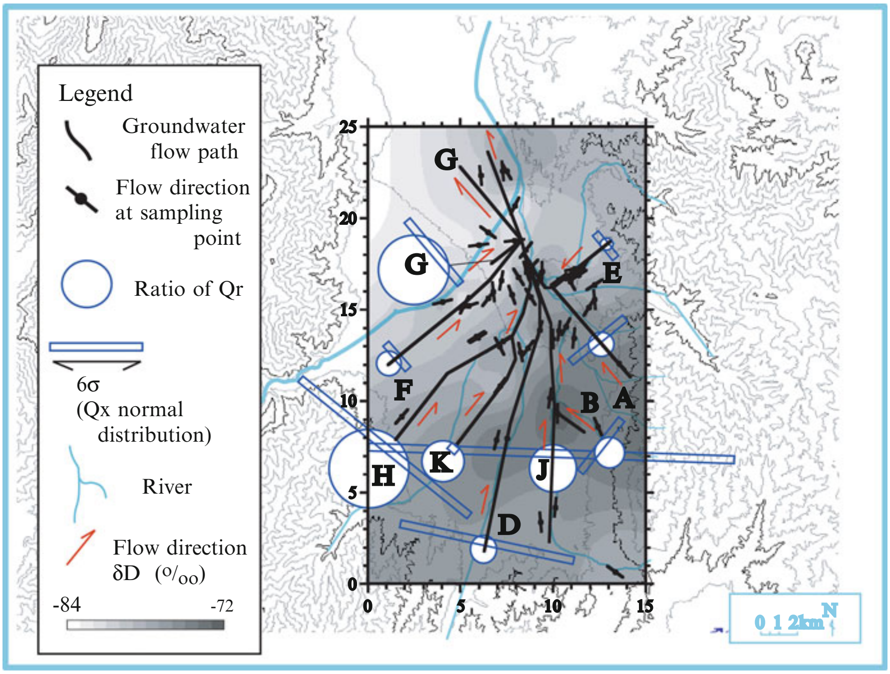

Fig. 6.14 Spatial distribution of flow paths (with Q_r and σ) inferred from the inversion analyses for the shallow groundwater system. (From Nakaya et al. 2007)

each flow path are set from the tentative end-members, and the initial values of Q_r and σ for all paths are set to 1.0 (m^3/s) and 1.0 (km), respectively, because data are not available for Q_r and σ. Figures 6.14 and 6.15 show the spatial distribution of flow paths (with Q_r and σ), inferred from the inversion analyses for the SGS and the DGS, respectively. The calculated δ^{18}O and δD inversion results for both groundwater systems agree well with the observed results, for a comparatively simple hypothesis (Figs. 6.16, 6.17). In this inversion, the residual vector between the observed and calculated ones in δ^{18}O is weighted by ten times to solve the normal equation to avoid the sensitivity difference between δ^{18}O and δD. The residual sums of squares [RSQ of Eq. (6.3)] between δ^{obs} and δ^{cal} of the final models are 57.8 (‰)^2 and 54.4 (‰)^2 in SGS and DGS, respectively. Tables 6.1 and 6.2 show the inverted parameters δ^{18}O, δD, Q_r, and σ for each path with estimated recharge altitudes in both groundwater systems, where Q_r are independent of each other in relationship to both SGS and DGS. The nine specific groundwater flow paths, inferred from the end-members in relationship to δ^{18}O and δD, are spatially separated from the multiflow system. Each path is concordant with the local flow directions, given the minimum of $\delta_{gradient}$ inferred from the observed δ^{18}O and δD data. Figure 6.18 shows the (δ^{18}O, δD) compositions of the rivers (mapped in Fig. 6.7), the flow paths, and end-members of SGS estimated from inversion analysis. The (δ^{18}O, δD) compositions of the rivers

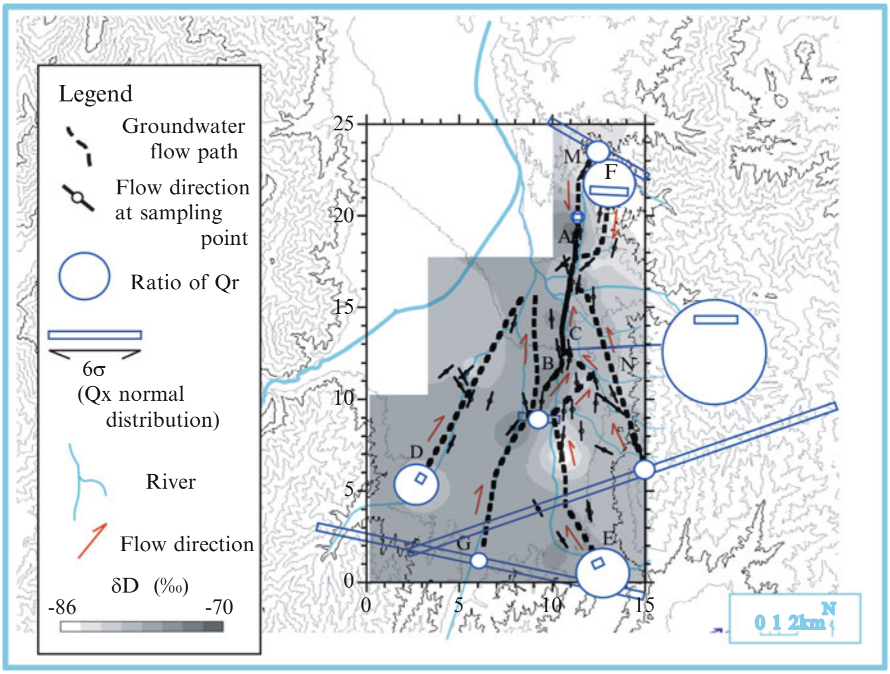

Fig. 6.15 Spatial distribution of flow paths (with Q_r and σ) inferred from the inversion analyses for the deep groundwater system. (From Nakaya et al. 2007)

are included in a nine-sided polygon composed of the estimated end-members in the SGS.

To check the accuracy of the final model, and to assess the sensitivity of the model, more inversion analyses are performed for the cases in which one or more end-members and flow path combinations are eliminated from the final model. These inversion cases of each end-member and flow path consist of two series: one inversion series for $\delta^{18}O$, δD, flow paths, Q_r, and σ, and another for $\delta^{18}O$, δD, Q_r, and σ holding the final flow paths, except paths eliminated from the final model. The results are shown in Tables 6.3 and 6.4 for the combinations G–F, H–K, and A–D in SGS and D–N and C–B in DGS, in which the differences in the isotopic compositions of the end-members or in the spatial distribution of flow paths are small (Fig. 6.13). These additional inversion results are higher than the final inversion results in RSQ, indicating that a more reliable model is composed of at least nine groundwater flow paths for each groundwater system.

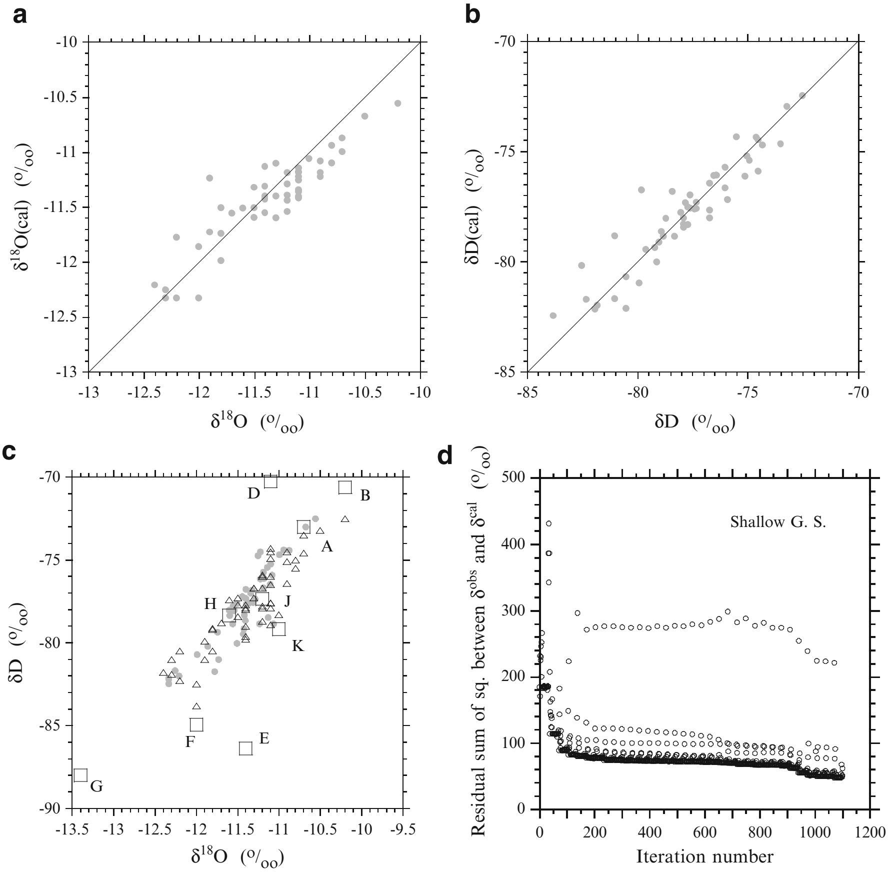

Fig. 6.16 Inversion results for δ^{18}O and δD in shallow groundwater. **a** Relationship between calculated δ^{18}O and the observed. **b** Relationship between calculated δD and the observed. **c** Relationship between δ^{18}O and δD. *Circles*, calculated; *triangles*, observed; *squares*, calculated end-member. **d** Relationship between iteration number in inversion and the residual sum of squares between the observed δ^{obs} and the calculated δ^{cal}. *G.S.*, groundwater system. (From Nakaya et al. 2007)

6.3.5 Discussion

The identification of flow paths and their extent, which had been obscure using spatial evaluations, could be estimated from the contour map of δ^{18}O and δD (as shown in Figs. 6.10 and 6.11). Considering δ^{18}O or δD as a natural tracer, these maps in Figs. 6.10a, b and 6.11a, b reflect the complexity of the groundwater flow systems, and consequently, the separation of the multiflow paths is not a simple task. The present method improves the procedure to find the groundwater flow paths and channels from field isotopic compositions, and gives rational

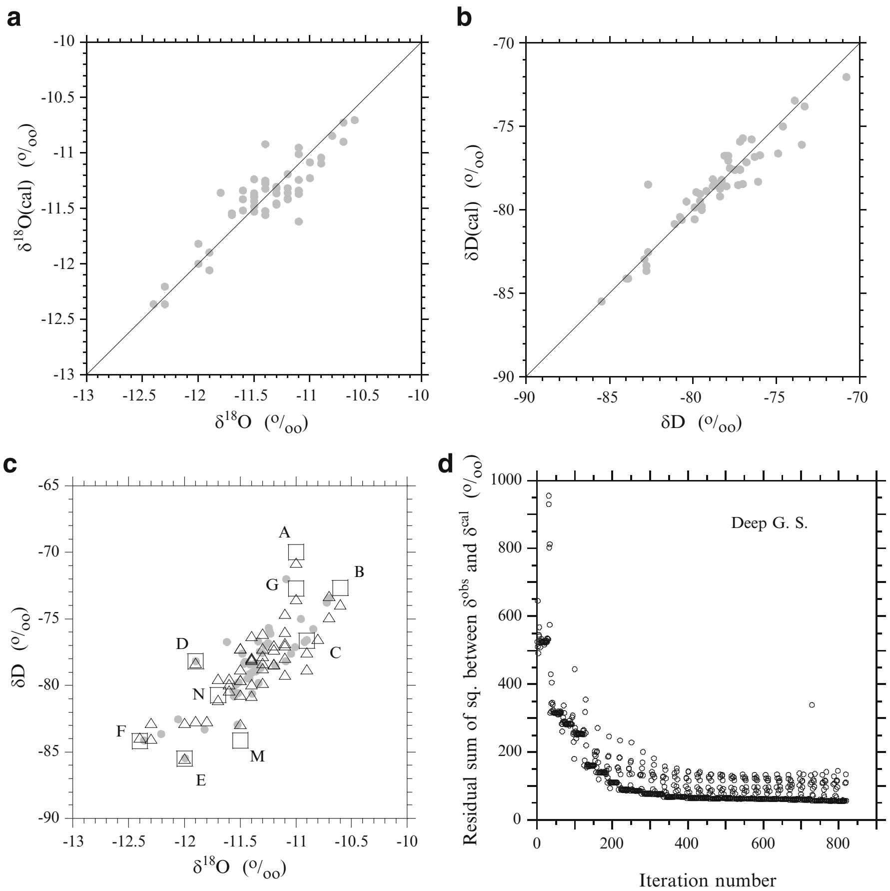

Fig. 6.17 Inversion results for $\delta^{18}O$ and δD in deep groundwater. **a** Relationship between calculated $\delta^{18}O$ and the observed. **b** Relationship between calculated δD and the observed. **c** Relationship between $\delta^{18}O$ and δD. *Circles*, calculated; *triangles*, observed; *squares*, calculated end-member. **d** Relationship between iteration number in inversion and the residual sum of squares between the observed δ^{obs} and the calculated δ^{cal}. (From Nakaya et al. 2007)

explanations and insights into the spatial distribution of the observed $\delta^{18}O$ and δD. Here, the present method is discussed and validated through the field application.

Komiya et al. (2003) reported that shallow and deep groundwater flow paths are spatially distinguishable using a contour map and the kinetic effect of $\delta^{18}O$ and δD in the same study area. Five flow paths were estimated for each of the shallow and deep groundwater systems. The inversion results clarify the nine groundwater flow paths for each, the shallow and deep groundwater systems, based on the consistent relationship between the ($\delta^{18}O$, δD) compositions and the spatial locations. The additional inversion results shown in Tables 6.3 and 6.4 also indicate that a model composed of at least nine groundwater flow paths for each groundwater systems is

Table 6.1 Inverted parameters for the shallow groundwater system

Flow path	$\delta^{18}O$ (‰)	δD (‰)	Q_r	σ (km)	Estimated recharge altitude (m) (avg)
G	−13.4	−88.0	1.90	0.773	1,850
F	−12.0	−84.9	0.211	0.330	1,400
H	−11.6	−78.4	2.17	2.02	1,250
K	−11.0	−79.2	0.684	0.0942	1,000
D	−11.1	−70.3	0.301	1.63	1,050
J	−11.2	−77.4	0.761	3.33	1,100
B	−10.2	−70.6	0.412	0.596	700
A	−10.7	−73.0	0.205	0.674	900
E	−11.4	−86.4	0.0587	0.285	1,150

Source: Modified from Nakaya et al. (2007)

Table 6.2 Inverted parameters in the deep groundwater system

Flow path	$\delta^{18}O$ (‰)	δD (‰)	Q_r	σ (km)	Estimated recharge altitude (m) (avg)
D	−11.9	−78.2	0.702	0.0772	1,350
G	−11.0	−72.7	0.102	2.98	1,000
B	−10.6	−72.7	0.143	0.367	850
E	−12.0	−85.5	1.03	0.0916	1,400
N	−11.7	−80.7	0.152	10.5	1,300
F	−12.4	−84.2	0.973	0.337	1,550
M	−11.5	−84.1	0.195	4.79	1,200
A	−11.0	−70.0	0.0586	0.0722	1,000
C	−10.9	−76.6	4.27	0.384	980

Source: Modified from Nakaya et al. (2007)

more accurate. The present model offers numerous significant insights on the groundwater flow in this study area, as follows.

The spatial patterns of groundwater flow paths illustrated in Figs. 6.14 and 6.15 clearly indicate differences in the locality of groundwater recharge zones and flow paths. Based on the inversion results performed in this study, the groundwater flow regimes can be viewed as follows. Groundwaters enter the basin in several recharge zones and subsequently mix during flow through the aquifers. The spatial patterns of the estimated groundwater flow paths are concordant with the groundwater potential contour and the directions of maximum hydraulic gradients in August 1993 (Figs. 6.10c, 6.11c) and the directions of the minimum gradient of δ in the shallow and deep groundwater systems, respectively. These results indicate that nine groundwater flows converge to the center of the basin from several recharge zones and then flow to the north in the SGS. Each flow path is likely to depend on the river network and fault system.

As shown in Fig. 6.18, the relationship between river and groundwater in isotopic compositions reflects the exchange of river water with groundwater flow paths, which indicates that the majority of river waters have the same sources as the

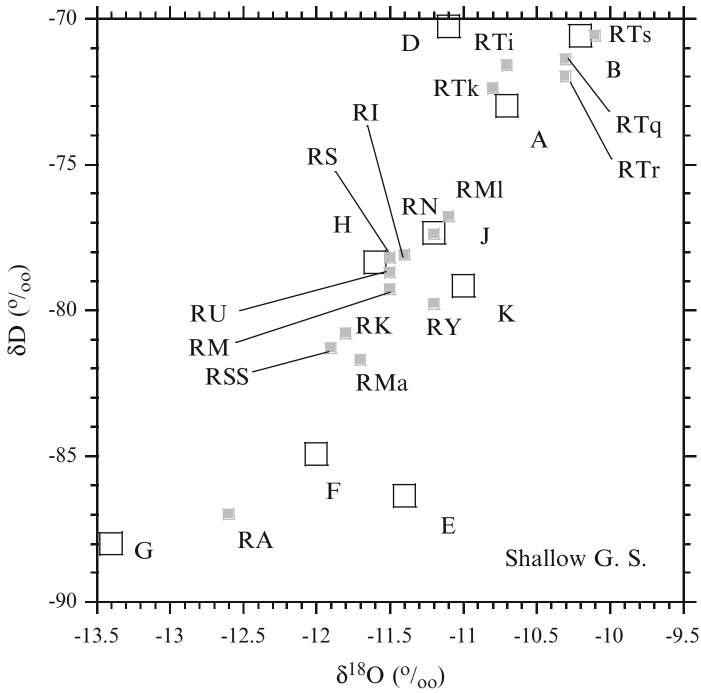

Fig. 6.18 The (δ^{18}O, δD) compositions of the river water (see Fig. 6.7) and those of the flow paths and end-members, A, B, D, E, F, G, H, J, and K of the shallow groundwater system estimated from inversion analysis. (From Nakaya et al. 2007)

Table 6.3 Results on sensitivity analysis of inversion in shallow groundwater system

Case	Parameters in inversion	RSQ (‰)^2	Problems
1 Delete path G	δ^{18}O, δD, path, Q_r, σ	84.3	Unnatural flow paths
2 Delete path G	δ^{18}O, δD, Q_r, σ	106	
3 Delete path K	δ^{18}O, δD, path, Q_r, σ	75.9	Unnatural flow paths
4 Delete path K	δ^{18}O, δD, Q_r, σ	69.8	
5 Delete path A	δ^{18}O, δD, path, Q_r, σ	73.9	Unnatural flow paths
6 Delete path A	δ^{18}O, δD, Q_r, σ	79.2	
7 Delete paths G, K and A	δ^{18}O, δD, path, Q_r, σ	98.9	Unnatural flow paths

RSQ residual sum of squares
Source: Nakaya et al. (2007)

Table 6.4 Results on sensitivity analysis of inversion in deep groundwater system

Case	Parameters in inversion	RSQ (‰)^2	Problems
1 Delete path C	δ^{18}O, δD, path, Q_r, σ	163	Unnatural flow paths
2 Delete path C	δ^{18}O, δD, Q_r, σ	63.6	
3 Delete path N	δ^{18}O, δD, path, Q_r, σ	214	Unnatural flow paths
4 Delete path N	δ^{18}O, δD, Q_r, σ	198	
5 Delete path D	δ^{18}O, δD, path, Q_r, σ	110	Unnatural flow paths
6 Delete path D	δ^{18}O, δD, Q_r, σ	72.8	

Source: Nakaya et al. (2007)

groundwaters. The model suggests that the majority of water from the river RA is composed of groundwaters with flow paths G and F in the SGS, because the (δ^{18}O, δD) composition of the river RA lies between the estimated (δ^{18}O, δD) compositions of flow paths G and F on the δ-diagram and the river RA is spatially close to flow paths G and F (Figs. 6.14, 6.18). Similar relationships with the (δ^{18}O, δD) compositions and spatial locations exist between the river RK and flow paths H and F, between the river RN and flow paths D and J, and between the river RY and flow paths E and A (and J) in the SGS.

Similarly, the majority of RT river water (at RTk, RTi, RTq, RTr, and RTs) is composed of groundwaters that follow flow paths B, D, A, and J in the SGS. The oxygen isotope shift may indicate the kinetic effect by evaporation resulting from the increased δ^{18}O composition of the waters at RTk, RTi, RTq, RTr, and RTs, downstream of the river RT. Especially, the (δ^{18}O, δD) composition at RTs of the river RT is coincident with the estimated high (δ^{18}O, δD) composition of flow path B, which contains the δ^{18}O at the surface at 600–800 m above sea level from the altitude effects of δ^{18}O shown in Fig. 6.9 (Waseda and Nakai 1983; Komiya et al. 2003). In addition, the RTs are spatially close to flow path B in the SGS. The path B groundwater may be a result of the kinetic effect caused by evaporation.

The (δ^{18}O, δD) compositions of rivers RS, RU, RM, and RI, which are surface flows above the shallow groundwater level from the eastern mountain of the basin, indicate a close resemblance to the composition of flow path J in the SGS. The compositional similarities and the spatial relationship between path J and these rivers indicate that they may be derived from the same source, that is, recharge water in the eastern watershed of the basin, which contains surface water at 1,400–1,600 m above sea level from the altitude effects of δ^{18}O (Komiya et al. 2003). Underflow of the river or riverbed water flow may play an important role in the SGS water resources. The concordance of the present model with the independent measurements from the inversion analysis already described, including the groundwater potential contours and the maximum hydraulic gradient directions, confirm the validity of the present approach in our understanding. The model would not work with lower end-members.

The current model also suggests that low δ^{18}O and δD fluids flow upward along fault zones and into the DGS (groups E and F in the DGS), whereas shallow groundwater, with the high δ^{18}O and δD of group B, flows into the deep aquifer once, then downward along faults and into the SGS and DGS along the footwall of a fault (Figs. 6.7, 6.13, 6.14, 6.15). The foregoing mechanism accurately describes the observations because the estimated (δ^{18}O, δD) compositions of flow paths B in both groundwater systems are high and concordant with each other and the fault is spatially situated between both paths B. In addition, the mixing between paths B and J in SGS may lower the (δ^{18}O, δD) composition of path B in DGS to the composition of path B in SGS. The potential contours in the DGS (Fig. 6.11c) support the intricate flows of estimated flow paths G, E, N, and B in the DGS. The flow path N can be controlled by a fault system. The flow path B in DGS appears to form the (δ^{18}O, δD) composition of path C by meeting and mixing with flow path E because path C is estimated to start from the meeting point of paths B and E spatially and the ($\delta^{18}O$, δD) composition of path C is estimated to lie between the estimated ones of flow paths B and E on the δ-diagram in the DGS.

The estimated parameters Q_r and σ in the SGS suggest that the recharge rate into the basin from the western and southwestern source areas is larger than that from the eastern source. This idea is consistent with the facts that the recharge areas of the western, southwestern, and eastern watersheds of the basin are about 559 km^2, 248 km^2, and 158 km^2, respectively, and the recharge waters in the south and southeastern watersheds of the basin flow into another basin. The groundwater flow regimes just described are consistent with the spatial variability within the major chemical composition (Fig. 6.8). The groundwater of flow path G in DGS reflects the spatial distribution of a low concentration in the southwestern part of the study region and is expected to be recharged at lower altitudes (approximately 700–900 m above sea level from the altitude effects of $\delta^{18}O$). High isotopic compositions are observed in the southwestern area, which is a short pathway from the recharge area.

Moreover, this method provides no quantitative groundwater flow model having flow velocities and pressure head. In the DGS, the isotope data in the northwestern part of the study area are not sampled because of a lack of wells; thus, the flow paths from the western source are never estimated. Therefore, the flow regime in the western part of the study area is impossible to analyze.

Additionally, because the flow rate and path of each groundwater flow path are estimated from the ($\delta^{18}O$, δD) compositions of sampling points in the summer and early autumn seasons, the groundwater flow regimes actually are likely to indicate those in the water-rich season and may reflect the differences in the ($\delta^{18}O$, δD) composition for the areas that are recharged at different times and through different flow paths in the system. If the present method was applied to the ($\delta^{18}O$, δD) compositions of samples in the dry season, the seasonal change in groundwater flow regimes could be seen.

Seasonal and annual changes in the flow rate and flow path of each groundwater flow path vary according to the recharge rate in the groundwater source area. However, in Central Japan, which includes this study area, the isotopic compositions of surface water are almost constant annually and indicate the average values of precipitation and the amount of rainfall at the source area (Kusakabe et al. 1970; Vogel and Vanurk 1975; Waseda and Nakai 1983). Waseda and Nakai (1983) reported that, in Central Japan, annual variation in the *d*-value of precipitation approximately ranges from 2 ‰ in summer to 25 ‰ in winter and that the variation in the *d*-value of surface waters ranges from 9.1 ‰ to 22.2 ‰ in the direction from the Pacific Ocean side (35.5°N) to the Sea of Japan (36.5°N). The variation in the *d*-value of groundwaters in this study area ranges from 10 ‰ to 17 ‰ in the summer and early autumn seasons (Fig. 6.12), indicating that major factors determining the isotopic composition of groundwaters in this area are the altitude effect and continental effect. The end-members in this study, which give the average values of recharge waters, should exhibit the isotopic compositions detected at the entrance to the sedimentary basin from the recharge area, which depend on a complex interaction of the kinetic effect of elevation, temperature, storm path, water–rock interaction, and diffusive recharge across large areas and the mixing of small sources. Variations in isotopic composition caused by kinetic effects accompanied by evaporation and precipitation along the pathway in the basin are not addressed by the model because of their minor effects in this study area, despite the impact of the kinetic effects on the major variations in the (δ^{18}O, δD) composition of recharged water in the source area. However, the path controlled by the kinetic effects in the basin can be separated as a distinct path when it appears on the δ-diagram as an end-member. In this study, we assume that all groundwaters are instantaneous mixes of the end members. Mixing between two or more waters strongly depends on the differences in their temperature and density. Because the small temperature differences were partly observed along path F in the DGS, their effects on the model should be reflected in the inverted σ of flow rate Q_x.

The present model must be improved in the near future using additional data to address the many problems described. However, it is possible to use differences in stable isotopic composition of waters to identify predominant flow paths in a complex basin where waters are recharged at various elevations and locations. Inversion analyses of stable isotope ratios similar to those performed in the present study are likely to prove very useful in future studies of groundwater flow regimes for the purpose of environmental preservation and management of water resources in urban areas.

6.4 Conclusions

A simple inversion method based on a simple mixing model has been proposed and demonstrated to spatially separate specific groundwater flow paths from the multiflow system, using stable isotopes of oxygen and hydrogen as natural tracers. Each groundwater flow is hypothesized to be a flow path with normal flow rate distribution in a direction perpendicular to the path. This method proposes that when more than one groundwater having different sources of (δ^{18}O, δD) merges through aquifers, the ratios of stable isotopes of oxygen and hydrogen at any point in the model can be calculated according to a simple mixing model. The number of groundwater flow paths and the (δ^{18}O, δD) of each path were inferred from the number of tentative end-members and the (δ^{18}O, δD) of each tentative end-member for the relationship between δ^{18}O and δD, respectively.

The following were found to be useful data for the spatial modeling of groundwater flow paths in the starting step of an inversion analysis using field data. (1) The local flow direction at each sample point, estimated using the directions given the minimum $\delta_{gradient}$, was inferred from the observed δ^{18}O and δD data; and (2) the spatial field distribution of the sample points were grouped around each end-member on the observed δ^{18}O versus δD graph, which were likely to be spatially distributed around or along each flow path.

The inversion resulting from the field data demonstrates that the specific groundwater flow paths inferred from the end-members of the relationship between δ^{18}O and δD can be spatially separated from the multiflow system. Each path obtained from the inversion was also concordant with the groundwater potential contour, the directions of maximum hydraulic gradients, and the local directions giving the minimum $\delta_{gradient}$ at the sample points. This inversion method clarifies the spatial separation of groundwater flow paths based on stable isotopes of oxygen and hydrogen when compared with the paths estimated using the spatial distribution in the contours of $\delta^{18}O$ and/or δD.

The groundwater flow regimes in the Matsumoto Basin estimated in this study are summarized next. Groundwaters enter the basin in several recharge zones and subsequently mix during flow through the aquifers, and each flow path is dependent on the river network and fault system. The estimated flow rate parameters in the SGS suggest that the recharge rate into the basin from the western and southwestern source areas is larger than that from the eastern source. The groundwater flow regimes estimated by inversion analysis are consistent with the spatial variability of the major chemical components.

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Chapter 7
Evaluation of Groundwater Vulnerability and Sustainability Using GIS

Tetsuya Hiyama, Insaf S. Babiker, and Mohamed A.A. Mohamed

Abstract Knowledge of spatial and temporal variability in groundwater quality is necessary to validate and compliment aquifer vulnerability estimates that have become important elements for sound resources planning. Here, to evaluate vulnerability of the unconfined groundwater, we assessed intrinsic aquifer vulnerability of the alluvial Nasuno Basin of Tochigi Prefecture, Japan, using the DRASTIC model. We also used a groundwater quality index, which synthesizes different available water quality data to delineate spatial variability in the overall groundwater quality. Data are mapped spatially in GIS (Geographic Information System) and the results integrated to assess the pollution risk and degree of sustainability of water quality in the basin. Although the study area was characterized by high to very high aquifer vulnerability, the groundwater quality was generally good with only limited zones showing relatively lower groundwater quality: the vicinity of the Naka and Houki Rivers and the lower part of the basin. This information clearly reflects the greater role of anthropogenic impacts (agricultural and urban activities) on the groundwater quality of the area.

Keywords Aquifer vulnerability • DRASTIC model • GIS • Groundwater quality index (GQI) • Sustainability of groundwater quality

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7.1 Introduction

Groundwater has been considered as an important source of water supply for many regions across the globe. It is usually of excellent quality. Being naturally filtered in its passage through the ground, it is often clear, colorless, free from microbial contamination, and therefore requires minimum treatment. However, groundwater quality is threatened by an ever-increasing number of soluble chemicals from urban, industrial, and agricultural activities. Monitoring and protecting the quality of groundwater is therefore, becoming a necessity to ensure safe water resources for human and animal use.

The chemistry (quality) of groundwater reflects inputs from the atmosphere, from soil and water–rock reactions (weathering), and from pollutant sources such as mining, agriculture, acid precipitation, and domestic and industrial wastes. The transport of contaminants from the point of application to the groundwater system is a function of the properties of the soil–rock strata above the aquifer and the type of pollutant (Melloul and Collin 1994). Accordingly, methodologies for assessing the degree of vulnerability of groundwater have been developed. Intrinsically, vulnerability is the property of the aquifer to receive and transmit contamination from anthropogenic sources. Much uncertainty is associated with aquifer vulnerability assessment as the result of insufficient representation of key factors such as soil media uncertainty regarding net recharge and hydraulic conductivity estimates (Evans and Myers 1990; Cavallin and Giuliano 1992; Rosen 1994), and lack of knowledge concerning the physical and chemical properties and attenuation processes of pollutants (Robins 2002), which require all vulnerability estimates to be validated by accurate field testing. Moreover, evaluating the contamination risk requires knowledge of the potential pollution sources or the actual condition of the underlying groundwater in addition to the intrinsic vulnerability.

Describing the overall water quality is difficult because of the spatial variability of multiple contaminants. Additionally, it is difficult to simplify to a few parameters. However, in the context of geo-indicators, five water quality indicators can be categorized: biological, physical, chemical, aesthetic, and radioactive. The range of indicators that can be measured is wide; therefore, the cost of a monitoring program to assess them all would be prohibitive. Consequently, vulnerability assessment can help delineate zones that are sensitive or under real risk of contamination. Resources can then be directed toward assessing important contaminants for the local environment in these specific zones.

In the current chapter, we demonstrate how integrating information on the physical environment that controls the transmission of contaminants in the groundwater system and on the actual condition of groundwater quality could provide useful knowledge for a sound resources management and land use planning. We first employ the DRASTIC model (Aller et al. 1987) of the United States Environmental Protection Agency (EPA) for assessing aquifer vulnerability in a selected case study area, the Nasuno Basin of Tochigi Prefecture, Japan. Second, we use the groundwater quality index (GQI) introduced by Babiker et al. (2007), which synthesizes different commonly available water quality data and provides a way to summarize the overall water quality. The GQI is used to validate the results of the aquifer vulnerability assessment. Finally, both the aquifer vulnerability and groundwater quality assessments are integrated to delineate regions that are under real risk of contamination and to evaluate the degree of sustainability of groundwater condition in the study area.

7.2 Study Area

The Nasuno Basin is one of the many alluvial basins in Japan. These basins are usually, shallow unconfined aquifers characterized by high groundwater storage, which are used mainly for domestic and agricultural purposes. They are also located in densely populated areas and therefore are subject to many contamination sources. In addition, previous studies (Hiyama and Suzuki 1991; Ohashi et al. 1994; Babiker et al. 2007) have reported a considerable degree of contamination in the groundwater of the Nasuno Basin, mainly by nitrate leaching from non-point agricultural sources.

Nasuno Basin is located on the northeastern part of Tochigi Prefecture, central Japan (Fig. 7.1). It represents a composite fan formed by the Naka, Kuma, Sabi, and Houki Rivers extending over an area of approximately 400 km^2. The altitude of the area ranges between 150 and 500 m above sea level (masl) with an average slope of 1/75 toward the south-southeast (SSE). The major cities are Otawara and Kuroiso with populations of 56,527 and 60,145 inhabitants, respectively. Irrigated rice cultivation is intensively practiced in the central and lower part of the fan and scattered stock farms are located mainly in the upper ranges of the fan. The mean annual precipitation and temperature are 1,426.5 mm and 12.4 °C, respectively, at Otawara City (mean of 22 years, 1987–2000). More precipitation occurs in the upper parts of the fan than in the lower parts, particularly between May and August.

As in most alluvial fans, the Nasuno Basin is thin in the upper northwest where there are ridges with an elevation of 500–320 masl. The thickness of the fan increases toward the southeast. Typically, the sand and gravels of the fan have high hydraulic conductivity ($3.3–9.4 \times 10^{-3}$ m/s; Choi 1976). Based on the field survey using 54 boreholes in the basin (Hiyama and Suzuki 1991), the depth from the surface to the groundwater table is generally shallow ranging from less than 1 m to 22.5 m. In the lower part of the basin the depth to groundwater is less variable, ranging between 5 and 10 m. Where the altitude is less than 260 masl, the depth to groundwater ranges between 0 and 5 m, with many springs representing typical discharge zones. Seasonal variation of water table depth ranges from a minimum of 0.07 m to a maximum of 6.72 m. The mean infiltration rate from the ground surface to the groundwater table is about 1 m/month (Choi 1976). Therefore, if we assume the median depth to the water table level is 12 m, the typical lag time for surface water to recharge groundwater is 1 year in the study area.

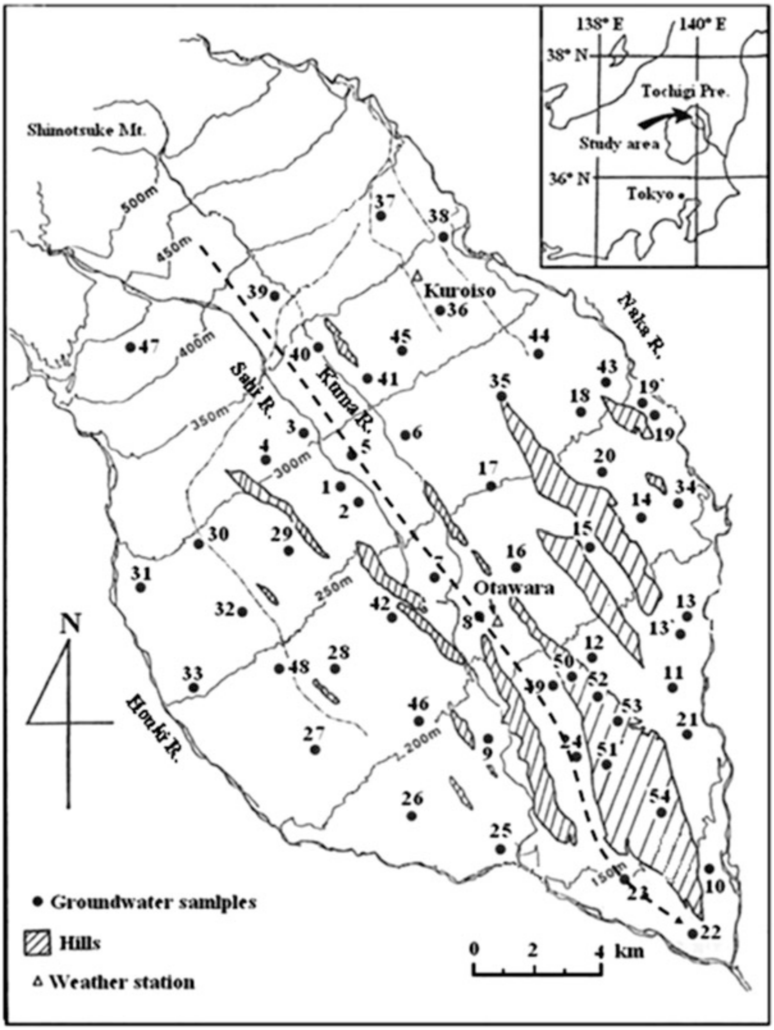

Fig. 7.1 Study area and location of groundwater samples. *Dashed arrow* indicates the general groundwater flow direction. (Modified from Babiker et al. 2007)

7.3 Materials and Methods

7.3.1 The DRASTIC Model

A DRASTIC model applied in a GIS environment was used to evaluate the vulnerability of the Nasuno aquifer. The DRASTIC model was developed by the U.S. Environmental Protection Agency (EPA) to evaluate groundwater pollution potential for the entire United States (Aller et al. 1987). It was based on the concept of the hydrogeological setting, which is defined as "a composite description of all the major geologic and hydrologic factors that affect and control the groundwater movement into, through and out of an area" (Aller et al. 1987). The acronym DRASTIC stands for the seven parameters used in the model, which are **D**epth to water, net **R**echarge, **A**quifer media, **S**oil media, **T**opography, **I**mpact of vadose zone, and hydraulic **C**onductivity. The model yields a numerical index that is derived from ratings and weights assigned to the seven model parameters. The various parameter classes are based on ranges specific to each parameter, which are then assigned a rating from 1 to 10 based on their relative effect on the aquifer vulnerability (Table 7.1). The seven parameters are then assigned weights ranging from 1 to 5, reflecting their relative importance to the overall intrinsic vulnerability. The DRASTIC Index is then computed applying a linear combination of all factors according to the following equation:

$$\text{DRASTIC Index} = D_rD_w + R_rR_w + A_rA_w + S_rS_w + T_rT_w + I_rI_w + C_rC_w \quad (7.1)$$

where D, R, A, S, T, I, and C are the seven parameters and the subscripts r and w are the corresponding rating and weights, respectively.

Several types of data were used to construct the thematic layers of the seven parameters of the DRASTIC model (Table 7.2). To capture the spatial variation in the hydrogeological setting in the Nasuno Basin, spatial analyses with GIS were conducted employing ILWIS, the GIS software of the International Institute for Geo-Information Science and Earth Observation (ITC). The datasets comprised six digital geologic maps, six digital soil maps, a digital elevation model (DEM), and six 1:50,000 topographic mapsheets showing the groundwater sample and weather station locations. Sample locations contained attribute data such as depth to water and groundwater quality. Features of the geologic and soil maps were digitized on screen to create point and segment maps of the different geographic entities. Point data from boreholes and weather stations were interpolated to create distribution maps of depth to water, precipitation, potential evapotranspiration, and hydraulic conductivity. The net recharge (mm/year) was defined as

$$\text{Net recharge (mm/year)} = \text{Potential recharge} \cdot \text{Recharge rate} \quad (7.2)$$

$$\begin{aligned}\text{Potential recharge (mm/year)} = {} & \text{Precipitation} \\ & - \text{potential evapotranspiration}\end{aligned} \quad (7.3)$$

The annual potential evapotranspiration was computed according to the Thornthwaite method (Thornthwaite 1948) using temperature and length of day from six weather stations distributed throughout the prefecture. The recharge rate was defined according to soil type using the rates provided originally by the DRASTIC model and using polynomial fitting function (Aller et al. 1987), so that the maximum soil rate allows 80 % of the potential recharge to percolate, although the minimum soil rate allows only 20 % to percolate.

Table 7.1 DRASTIC ratings and the corresponding ranges for the seven parameters (Aller et al. 1987)

	D (5)	**R (4)**	**T (1)**	**C (3)**	**A (3)**		**S (2)**		**I (5)**	
Rating	Range (m)	Range (mm/year)	Range (slope %)	Range (m/s)	Range	Rating	Range	Rating	Range	Rating
1	>100	0–50.8	>18	$<4.7 \times 10^{-5}$	Massive shale	2 (1–3)	Non-shrinking and non aggregated clay	1	Confining layer	1
2	75–100	–	–	4.7×10^{-5}–1.4×10^{-4}	Metamorphic/igneous	3 (2–5)	Muck	2	Silt/clay	3 (2–6)
3	50–75	50.8–101.6	12–18	–	Weathered metamorphic/igneous	4 (3–4)	Clay loam	3	Shale	3 (2–5)
4	–	–	–	1.4×10^{-4}–3.3×10^{-4}	Glacial till	5 (4–6)	Silty loam	4	Metamorphic/igneous	4 (2–8)
5	30⁻50	–	6–12	–	Bedded sandstone, limestone and shale sequences	6 (5–9)	Loam	5	Limestone	6 (2–7)
6	–	101.6–177.8	–	3.3×10^{-4}–4.7×10^{-4}	Massive sandstone	6 (4–9)	Sandy loam	6	Sandstone	6 (4–8)
7	15–30	–	–	–	Massive limestone	6 (4–9)	Shrinking and/or aggregated clay	7	Bedded limestone, sandstone and shale	6 (4–8)
8	–	177.8–254	–	4.7×10^{-4}–9.4×10^{-4}	Sand and gravel	8 (4–9)	Peat	8	Sand and gravel with significant silt and clay	6 (4–8)
9	5–15	>254	2–6	–	Basalt	9 (2–10)	Sand	9	Sand and gravel	8 (6–9)
10	0–5	–	0–2	$>9.4 \times 10^{-4}$	Karstic limestone	10 (9–10)	Gravel	10	Basalt	9 (2–10)
							Thin or absent	10	Karstic limestone	10 (8–10)

The shaded cells are the DRASTIC ranges and media types present in the study area

D, depth to water; R, net recharge; A, aquifer media; S, soil media; T, topography; I, impact of vadose zone; C, hydraulic conductivity

Numbers in bold and between brackets are the weights assigned to emphasize the importance of each parameter

Table 7.2 Data used to construct the individual layers for the DRASTIC and the groundwater quality (GQI) models

Data type	Source	Format	Scale	Date	Used for
Seasonal borehole data (chemistry and water table level)	Authors	Tables (4) Location maps (6 maps)	1:50,000	1989–1991 1983	GQI Depth to water
Climatic data (annual mean temperature and precipitation)	Japan Meteorological Agency website: (http://data.kishou.go.jp)	Table (6 stations)		1979–2000	Net recharge
Geology map	Ministry of Land, Infrastructure and Transport website: (http://tochi.milt.go.jp)	Digital maps	1:50,000	1987–1994	Aquifer and vadose zone media
Soil map	Ministry of Land, Infrastructure and Transport website: (http://tochi.milt.go.jp)	Digital maps	1:50,000	1987–1994	Soil media and net recharge
Topographic data (DEM)	Geographical Survey Institute of the Ministry of Land, Infrastructure and Transport	Digital on CD-ROM (24 maps)	50-m grid size	2003	Topography
Hydraulic conductivity	Groundwater level annual report No 26, Tochigi Prefecture Living Environment Section, Environmental Bureau, Environmental Control Section, and literature	Point data (3 points) (1 point)		2004 1976	Hydraulic conductivity

Table 7.3 Statistics of seven groundwater quality parameters measured in "spring" from the Nasuno Basin and the corresponding maximum threshold values according to the WHO

Parameter (mg/l)	Mean (±SD)	WHO threshold value (mg/l)
Ca^{2+}	13.4 (4.7)	300
Mg^{2+}	4.6 (1.6)	300
Na^{+}	7.6 (2.9)	200
Cl^{-}	8.7 (2.8)	200
NO_3^{-}	22.0 (11.2)	50*
SO_4^{2-}	20.8 (7.1)	250
TDS	88.9 (25.9)	600

* Represents a guideline value assigned by the WHO for the nitrate since it might inflict potential health risk.

The topography (in slope %) was computed from the DEM according to the following equation:

$$\text{Topography} = \sqrt{}(\text{DX})^2 + (\text{DY})^2/\text{Pixel size} \cdot 100 \tag{7.4}$$

where DX and DY are the horizontal and vertical gradients.

7.3.2 The Groundwater Quality Index (GQI)

To evaluate the overall groundwater quality in the study area, the GQI recently introduced by Babiker et al. (2007) was used. Seasonal (four seasons: spring, summer, autumn, winter) groundwater quality data collected from the Nasuno Basin from more than 50 water wells were utilized. The locations of groundwater samples are shown in Fig. 7.1. The data set includes measurements of several physical and chemical parameters of groundwater obtained by in situ measurements and laboratory analytical techniques. Seven parameters that are listed in the World Health Organization (WHO) (2004) guidelines for drinking water quality were selected from the data set to generate the GQI. Standards for drinking water were chosen because human health is taken as priority; also, the high quality required for drinking water makes it suitable for many other purposes. Six parameters (Cl^{-}, Na^{+}, Ca^{2+}, Mg^{2+}, SO_4^{2-}, and total dissolved solids, TDS) are in the category of chemically derived contaminants that could alter the water taste, odor, or appearance and affect its "acceptability" by consumers (WHO 2004). Fixed guidelines have not been established for these chemicals; thresholds for maximum desired concentrations only were discussed. One parameter (NO_3^{-}) was listed under the category of chemicals that might inflict "potential health risk" and was assigned a guideline value (50 mg/l; WHO 2004).

For the purpose of implementing the GQI, only the spring season data were selected (Table 7.3). A significant change in water quality, particularly the increase of nitrate concentration in groundwater, was observed during spring (Hiyama and Suzuki 1991; Babiker et al. 2007). This change was attributed to the increase of input of nitrate with irrigation water resulting from the commencement of the season of rice cultivation. Data from all seasons are used to address the temporal variation of groundwater quality in the basin.

The GQI is generated in four steps: first, a concentration map referred to as "primary map I" was constructed for each parameter from the point data using Kriging interpolation. Following interpolation, a raster map with 50-m pixel size was generated. Second, to relate the groundwater quality data to a universal norm, the measured concentration, X', of every pixel in each "primary map I" was related to its desired WHO standard value (Table 7.3), X, using a normalized difference index:

$$C = \left(X' - X\right)/\left(X' + X\right) \tag{7.5}$$

The resultant "primary map II" thus displays for each pixel a contamination index value ranging between −1 and 1. Third, the primary map II was then ranked between 1 and 10 to generate a "rank map" for each parameter using the following expression:

$$r = 0.5\,C^2 + 4.5\,C + 5 \tag{7.6}$$

where C stands for the contamination index value for each pixel determined from primary map II, and r stands for the corresponding rank value. A rank of 1 indicates minimum impact on groundwater quality; the rank of 10 indicates maximum impact. Use of this expression results in the minimum contamination index level (−1) being set equal to 1, the median level (0) set equal to 5, and the maximum level (1) set equal to 10. Finally, the GQI was calculated as follows:

$$\mathrm{GQI} = 100 - ((r_1 w_1 + r_2 w_2 + \cdots + r_n w_n)/N) \tag{7.7}$$

where r is the rank (1–10) of the parameter, w is the relative weight of the parameter, and N is the total number of parameters ($N = 7$) used in the suitability analyses.

The GQI represents an averaged linear combination of factors. The weight (w) assigned to each parameter indicates its relative importance to groundwater quality, and corresponds to the mean rank (mean r) determined from its respective "rank map" (1–10). Parameters that have a higher impact on groundwater quality (high mean rank) are assumed to be more important in evaluating the overall groundwater quality. Particular emphasis was given to contaminants that pose potential risk to human health (e.g., nitrate); thereby, w was determined from the "mean $r + 2$" ($r \leq 8$). Dividing by the total number of parameters averages the data and limits the index values between 1 and 100. In this way, the impact of individual parameters is greatly reduced and the index computation is never limited to a certain number of chemical parameters.

7.4 Results and Discussion

7.4.1 The DRASTIC Parameters and Aquifer Vulnerability

Table 7.4 summarizes the dominant ranges and media types of the DRASTIC parameters in the study area and their corresponding rates. Based on the foregoing rankings for each DRASTIC parameter, the aquifer vulnerability of the Nasuno Basin fell under the "High" and "Very High" classes (index value ranges between 140–179, and 180–199, respectively) (Fig. 7.2). The very high vulnerability regions in the vicinity of the Naka River, between the Sabi and Kuma Rivers and at the southeastern end of the fan, can be attributed to a combination of shallow groundwater, medium texture of soil, and vadose zone media (sandy loam and gravel sand, respectively) and high hydraulic conductivity.

7.4.2 The Groundwater Quality

The groundwater quality of the Nasuno Basin is generally high (mean GQI = 95.9, SD = 0.5, maximum quality = 100). The GQIs are classified based on a fixed interval of area percentage in the study area. Nine groundwater quality classes are identified in the study area at a 10 % interval. The classification scale reflects the same level of detail for the eight lowest classes corresponding to 80 % of the total area. The 20 % of the study area of the best groundwater quality was considered less important and was assigned a single class interval (>80 %) (Fig. 7.3).

Two gradients of groundwater quality can be observed in the study area. First, there is a decrease in groundwater quality from northwest to southeast following the general groundwater flow direction. In a recent study of the nitrate contamination of groundwater in the Nasuno Basin (Babiker et al. 2007), an increasing trend of nitrate concentration was reported along groundwater flow direction from northwest to southeast; this is associated with the increase in density of irrigated rice fields. Also, Ohashi et al. (1994) concluded from a study involving nitrate isotopes that the groundwater nitrate in the middle part of the Nasuno Basin originates

Table 7.4 Summary of the final DRASTIC values in the study area

DRASTIC parameters	Ranges and media types	Rating
Depth to water	1.5–18.5 m	6–9
Net recharge	>250 mm/year	8–9
Aquifer media	Sand and gravel	8
Soil media	Sand loam and silt loam	6 and 4
Topography	1–2 %	10
Impact of vadose zone	Loam and semi-consolidated gravel, sand, and mud	6–8
Hydraulic conductivity	4.7×10^{-4}–5.5×10^{-3} m/s	8–10

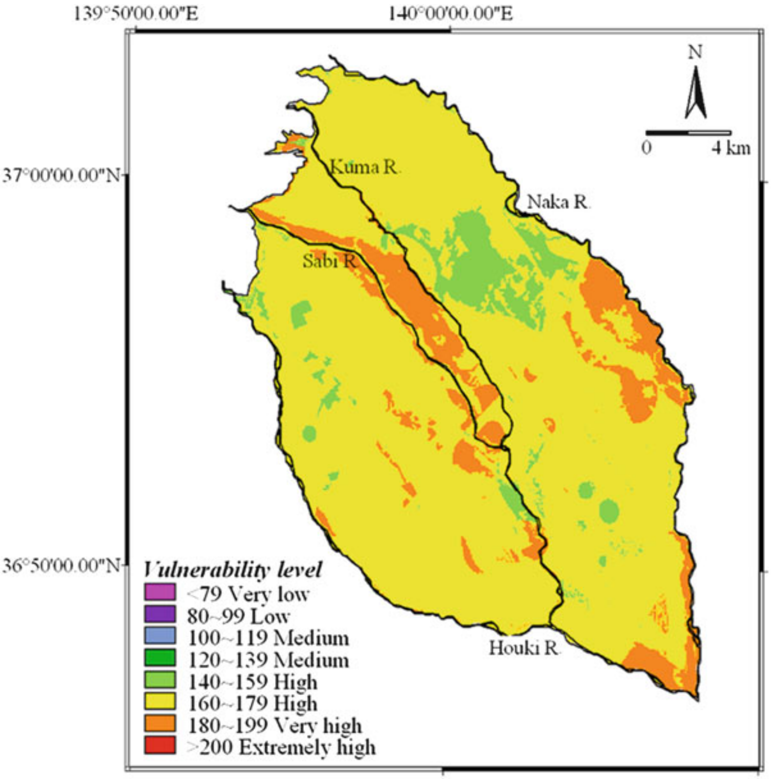

Fig. 7.2 Intrinsic aquifer vulnerability map of the Nasuno Basin

mainly from animal waste whereas that in the lower part of the basin originates from chemical fertilizers, clearly reflecting the effect of contamination transport by groundwater flow. In addition, the upper northwest part of Nasuno Basin is characterized by a very low population density, which lowers the chance of anthropogenic contamination. In contrast, the lower part of the basin is densely populated.

The second gradient is a decrease of groundwater quality from the central line of the basin toward the Naka River and Houki River borders. Hiyama and Suzuki (1991) have identified a similar variation pattern in groundwater quality and suggested five regions of different qualities. These groundwater quality regions seem to delineate the structural and geomorphological differences among the old terraces originally formed by the fan deposit. Nevertheless, other mechanisms might also be responsible for this spatial variation of groundwater quality in the Nasuno Basin. The Houki River (southwest border), which drains the densely populated area of Nasuno Hot Spring in the north, represents a potential source of recharge to the groundwater. Therefore, urban pollutants may be introduced into the groundwater system and transported down gradient from this area. Wakui and Yamanaka (2006) applied stable isotopes of oxygen and hydrogen as natural tracers and found that the groundwater near the Houki River is predominantly recharged by

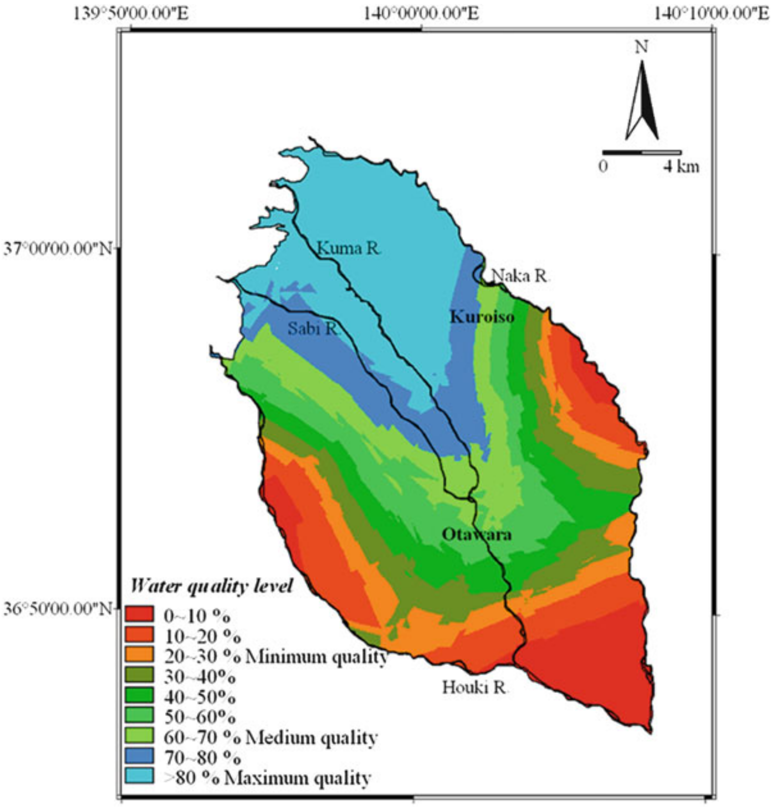

Fig. 7.3 Map of groundwater quality level in the Nasuno Basin. (Modified from Babiker et al. 2007)

precipitation where heavy rain tends to result in the rapid rise in the water table. They thereby pointed out that increases of the nitrate concentrations were through a piston-type flow of water. Near the Naka River (northeast border) the low groundwater quality might be attributed to the relatively high population density in that region (Kuroiso City), where precipitation is the only source of groundwater recharge (Wakui and Yamanaka 2006).

7.4.3 Validation of Aquifer Vulnerability and the Pollution Risk Assessment

Civita (1994) proposed that the “integrated vulnerability,” the equivalent of contamination (pollution) risk, could be obtained by overlying a representation of the actual pollution sources on the intrinsic vulnerability. The cross-overlay

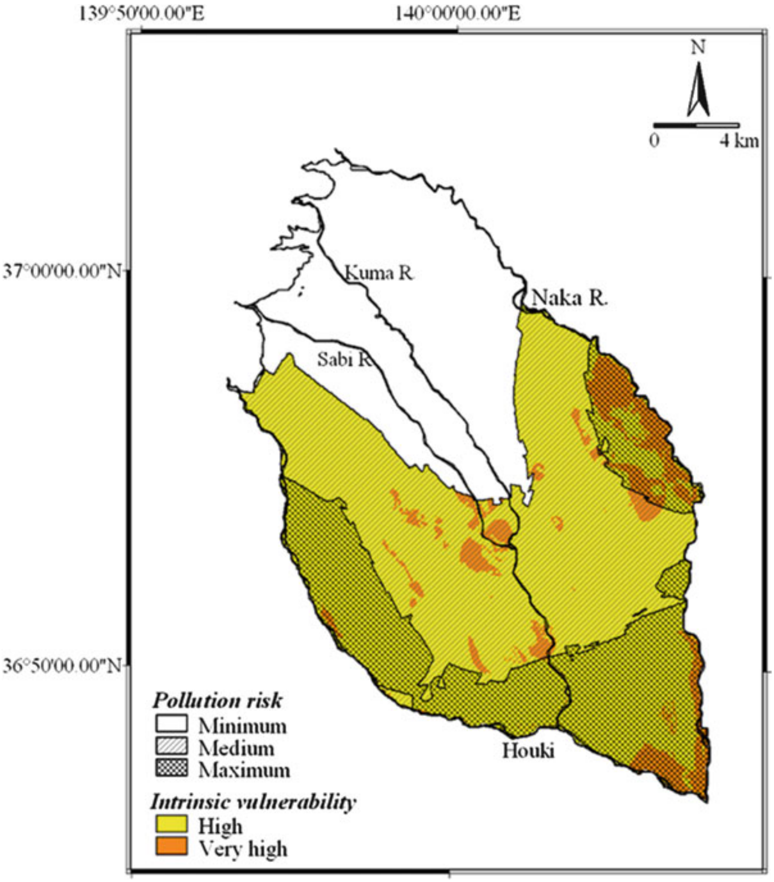

Fig. 7.4 Pollution risk map of the Nasuno Basin integrating intrinsic aquifer vulnerability and groundwater quality

operation of the GIS was therefore, used to validate the results of aquifer vulnerability assessment using the groundwater quality model results. This approach spatially relates the degree of vulnerability with the underlying groundwater quality level. The cross-operation performs an overlay of two raster maps by combining pixels at the same locations in both maps and tracking all the combinations that occur between the different values or classes in both maps (ITC-ILWIS 2001). The proportion of aquifer vulnerability class unit area that spatially overlaps with each groundwater quality level is obtained. A simple multiple map overlay was then applied to visually display the relationship between aquifer vulnerability and underlying groundwater quality (Fig. 7.4), such that the vulnerability classes could be critically considered only where a groundwater pollution risk exists.

Most of the study area is characterized by "high" aquifer vulnerability. The largest proportion of land area in the three groundwater quality levels overlapped spatially with the area of high aquifer vulnerability (83 %, 93 %, and 87 % of the

Table 7.5 Proportion of aquifer vulnerability level area overlapping spatially with each groundwater quality level

Groundwater quality level	Total area (km^2)	Aquifer vulnerability level	
		High (%)	Very high (%)
Minimum	113.1	83	17
Medium	162.5	93	7
Maximum	127.9	87	13

minimum, medium, and maximum quality levels, respectively) (Table 7.5). Regions characterized by "very high" aquifer vulnerability might be expected to display a relatively lower degree of groundwater quality. However, these very high vulnerability areas overlap spatially with only a small percentage of the area (17 %, 7 %, and 13 % of the minimum, medium, and maximum quality levels, respectively) (Table 7.5). The disagreement in spatial pattern between the degree of aquifer vulnerability and the level of groundwater quality in most parts of the area clearly reflects a greater role of anthropogenic impacts on the groundwater resources of the Nasuno Basin. The map-overlay results indicate that there is a good spatial overlap between the "very high" aquifer vulnerability and the minimum groundwater quality only in the vicinity of the Naka River and the southeastern end of the basin, indicating that these regions are already under high risk of groundwater pollution attributed mainly to anthropogenic impacts (land use and population density). The elongated zone of "very high" aquifer vulnerability around the central line of the basin is partly associated with medium groundwater quality, indicating moderate contamination. In the upper parts of the basin, there is limited risk of pollution, as indicated by the high quality of the underlying groundwater. Generally, although most of the study area is characterized by high vulnerability, the lower part of the fan and the vicinity of the Houki and Naka Rivers seem to be under high risk of pollution, as indicated by the low quality of the underlying groundwater.

7.4.4 Sustainability of Groundwater Quality

Sustainability of groundwater conditions is assumed to rely on the degree of aquifer vulnerability and a measure of temporal variation of groundwater quality. In a groundwater system characterized by high infiltration rate and high hydraulic conductivity, such as the Nasuno Basin, the seasonal variation of groundwater quality is significant. The multi-seasonal data set allows estimation of the degree of seasonal variation of groundwater quality in the Nasuno area. The coefficient of variation, which is a measure of variability in time, can be expressed as

$$V = ((\text{standard deviation}/\text{mean}) \cdot 100) \tag{7.8}$$

where V was calculated for each groundwater quality parameter in boreholes that were sampled at least three seasons a year. The total variation in each borehole was

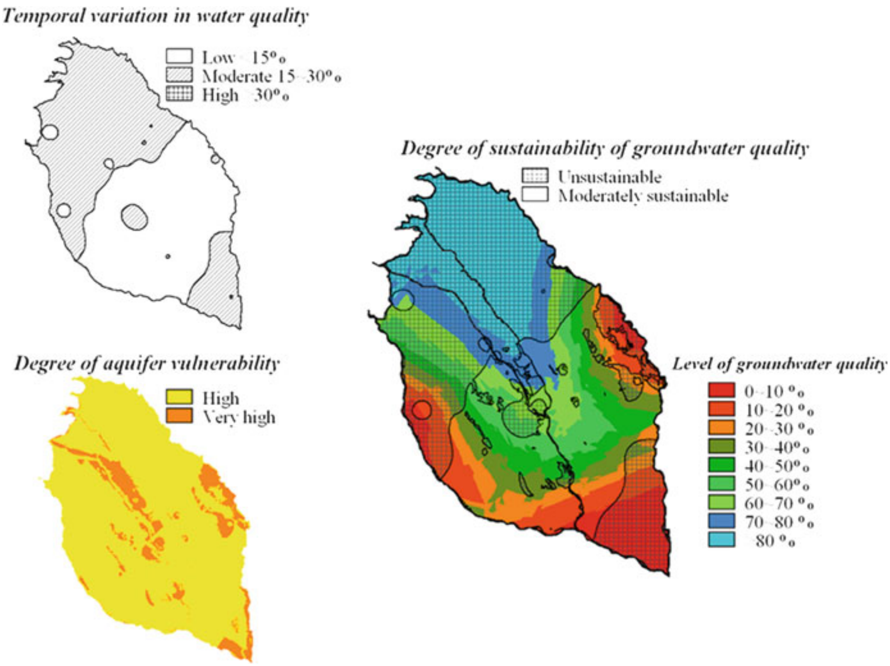

Fig. 7.5 Integration of temporal variability of groundwater quality, aquifer vulnerability, and level of groundwater quality (GQI) to evaluate the degree of sustainability of the current groundwater quality in the Nasuno Basin. (Modified from Babiker et al. 2007)

then calculated as the sum of variations of the different groundwater quality parameters.

$$V_{\text{total}} = V_1 + V_2 + \cdots V_n \tag{7.9}$$

where n is the number of groundwater quality parameters. A seasonal variation map was generated from the borehole point data using moving-average interpolation. The seasonal variation map was then integrated with the aquifer vulnerability map and the GQI map (Fig. 7.5); this results in the degree of sustainability of groundwater quality increasing when both aquifer vulnerability and temporal variation of water quality decrease.

Groundwater quality was more variable in the upper and lower parts of the basin (V, 15–30 %), likely as a result of the seasonality of precipitation and irrigation (Fig. 7.5). Integration of the aquifer vulnerability and variation index indicates that in the upper and lower parts of the basin groundwater quality is unsustainable because of the moderate temporal variability and the high pollution potential of the aquifer (Fig. 7.5).

7.5 Summary and Conclusion

In this chapter, GIS was employed for capturing, analyzing, and modeling different environmental data to evaluate aquifer vulnerability and groundwater quality in the Nasuno Basin, Tochigi Prefecture, Japan. The groundwater quality was used to validate and complement the aquifer vulnerability assessment and to derive an assessment of the actual pollution risk and the sustainability of groundwater quality in the study area.

The study area displays high to very high aquifer vulnerability, as dictated by the natural characteristics of the hydrogeological setting. The aquifer was particularly vulnerable in the vicinity of the Naka River at the northeastern edge of the basin and in a central zone between the Sabi and Kuma Rivers. The groundwater of the basin is generally of high quality. Two distinct gradients in groundwater quality were observed. One is a decrease in groundwater quality from the northwest to the southeast following the general groundwater flow direction, which is attributed mainly to the decrease in depth to groundwater table and the increase of pollutants (mainly nitrate) input from chemical fertilizers. The other gradient is a decrease of groundwater quality from the central line of the basin toward the northeast (Naka River) and southwest (Houki River) borders of the basin, which can be attributed mainly to structural and geomorphological differences in the basin.

Validation of the aquifer vulnerability assessment allowed recognition of the degree of agreement between the estimated aquifer sensitivity to pollution and the actual condition of the underlying groundwater. This approach helps to identify zones that are under "real" risk of pollution from anthropogenic sources. Although the study area is characterized by high to very high aquifer vulnerability, the groundwater quality was generally good, with only limited zones showing relatively lower groundwater quality: the vicinity of the Naka and Houki Rivers and the lower southeastern part of the basin. This observation clearly reflects the greater role of anthropogenic impacts (agricultural and urban activities) on the groundwater quality of the area. Nevertheless, assessment of the degree of sustainability of groundwater quality indicated that in the upper and lower parts of the basin the current groundwater quality is provisional and subject to change under varying conditions as a consequence of the moderate temporal variability and high pollution potential of the aquifer.

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Chapter 8
The Kabu-ido System: Implications for Current Groundwater Management Policy

Takahiro Endo

Abstract The Kabu-ido system was a customary groundwater management practice that was once used in the southern part of Noubi Plain in the Tokai area of Japan. The system had two features: limitation of the total number of wells and groundwater users' obligation to pay economic compensation. From the theoretical point of view, it is an example of the Coase theorem, because it can be regarded as an institution by which the stakeholders internalized negative externalities caused by groundwater pumping through private negotiations. Although the Kabu-ido system is not used any more, it still contains implications for current groundwater management policy including the importance of third-party monitoring, usefulness of issue linkage in the process of conflict management, effectiveness of economic tools on the use of groundwater pumping restrictions, and significance of an ear-marked tax.

Keywords Externality • Groundwater • Kabu-ido • Pump tax • Coase theorem

8.1 Introduction

The purpose of this chapter is to introduce a groundwater management system (the Kabu-ido system) that once was used in the southern part of Noubi Plain in the Tokai area of Japan. The implications of the Kabu-ido system for current groundwater management are then discussed.

The southern part of the Noubi Plain in the Tokai area of Japan is a low-lying area composed of a large delta that is commonly subjected to severe flooding.

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Local residents in this region have developed a unique system known as the ring-levee (Waju in Japanese) to protect against flooding. However, they sometimes faced severe water shortage in years during which there was low precipitation. To address these occasional water shortages, irrigation by artesian wells expanded rapidly in the ring-levee area from the early to mid-nineteenth century. Although the development of artesian well systems greatly stabilized the water supply within the ring-levees, it led to accumulation of drainage water in the lower part of these areas.

The Kabu-ido system was developed to address conflicts related to drainage within the ring-levee systems. It was established in the nineteenth century and used until around 1905 to solve drainage problems and provide a method for regulation of uncoordinated groundwater pumping (Katano 1941, p. 68). In Japanese, "Kabu" means "privilege to do a business" (Miyamoto 1977, p. 59) and "ido" means "well"; therefore, "Kabu-ido" can be interpreted as "privileged well" or "the special right to dig wells." Accordingly, considering the Kabu-ido systems and their development can provide useful lessons for current groundwater management systems.

The Kabu-ido system has been investigated since the 1930s. Bekki (1932) and Nakazawa et al. (1936) introduced the Kabu-ido system as a part of their investigations of ring-levee systems. Katano (1941), Mori (1964) and Matsubara (1968) later clarified the Kabu-ido systems in the Takasu and Fukuzuka ring-levees based on a wide review of historical documents. The Kabu-ido system has also been described in history books published by local governments in ring-levee areas such as Kaizu, Hirata, and Wanouchi (Kaizu 1970a, b; Mori 1964; Editorial Committee of a History of Wanouchi Town 1981; Wanouchi Town 1991). These studies were followed by recent works by Andoh (1975), Ito and Aoki (1987), and Itoh (2002).

Although these previous studies clarified the Kabu-ido system, they only focused on the system itself and did not consider the implications that these systems have on current groundwater management policies. Accordingly, this study was conducted to investigate the Kabu-ido system from the aspect of policy science.

8.2 The Kabu-ido System

8.2.1 Description of the Kabu-ido System

The Kabu-ido system was created in the ring-levee area at the confluence of the Ibi, Nagara, and Kiso Rivers (Fig. 8.1). Ring-levees were developed in this area for several reasons. (1) The area was prone to natural flooding. (2) The central government at the time constructed a 50-km levee on the east side of the Kiso River for military purposes. Because of this levee, floods concentrated on the west side of Kiso river. (3) This area was ruled by a number of landlords in a fragmented fashion that prevented implementation of effective flood controls. Accordingly, the local residents enclosed themselves in series of levees to protect themselves (Bekki 1932, p. 234, pp. 248–250; Nakazawa et al. 1936, pp. 15–16). Ring-levees are still present

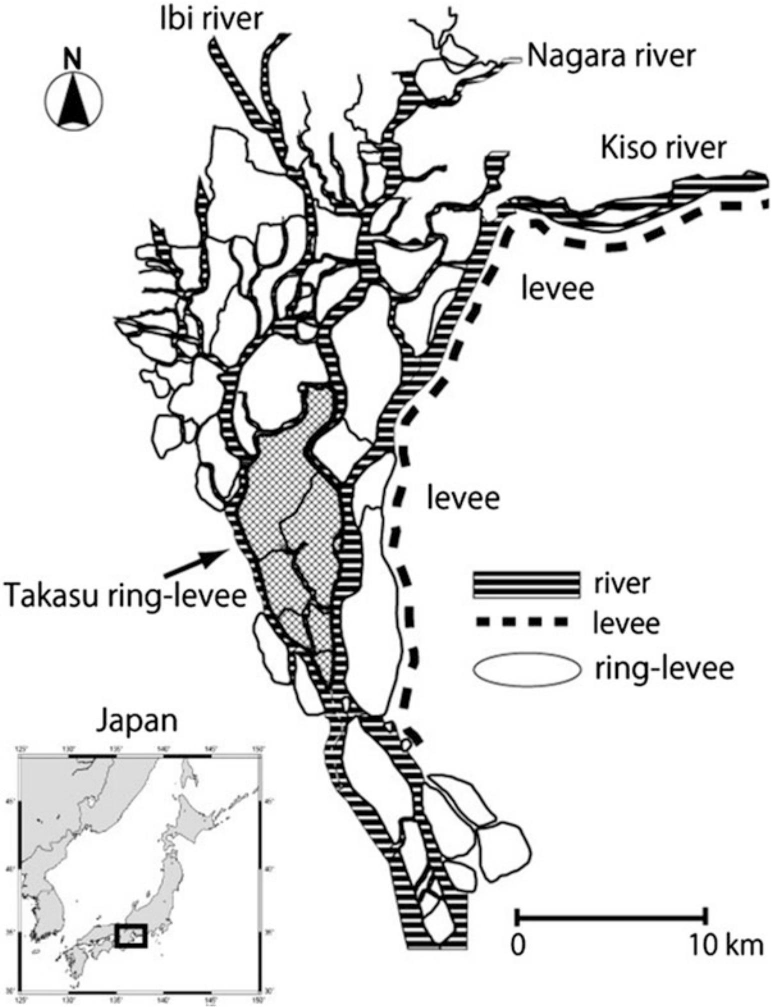

Fig. 8.1 Ring-levees around 1870 (based on a map made by Yasuo Itoh in Nanno Town 1978)

in this region, forming a reverse triangle that extends 40 km from east to west and 50 km from north to south and encompassing a total of 1,800 km^2 (Ito 2001, p. 3).

The Kabu-ido systems previously existed in Takasu, Fukuzuka, and Tagi ring-levees (Katano 1941, p. 56; Yourou Town 1978, p. 294). This study investigated the Kabu-ido system in the Takasu ring-levee (Fig. 8.2), for which the most abundant documents are available. Although residents inside the Takasu ring-levee faced occasional flooding, they also suffered water shortages in years of less precipitation. To address this issue, artesian wells were implemented in the nineteenth century (Mori 1964, p. 918; Matsubara 1968, p. 494). Although the difference in altitude is very small, the northern part of the Takasu ring-levee is higher than the southern

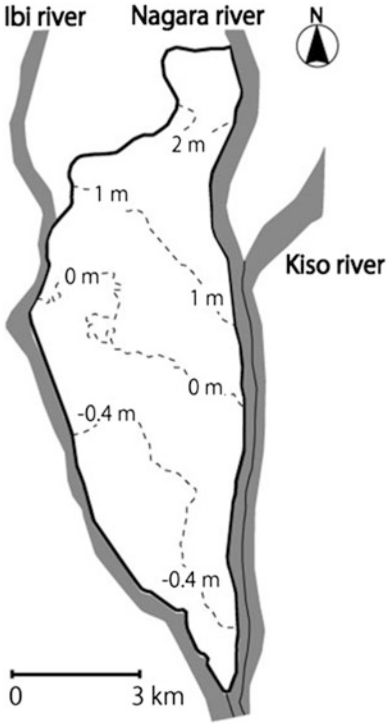

Fig. 8.2 Takasu ring-levee and its altitude (based on a figure provided by the Kaizu Educational Committee (2009), p. 20)

part. Artesian wells were primarily constructed in villages in the northern part of the town, where the altitude was more than 1 m above sea level (hereinafter referred to as the upper villages). Because of geographic conditions the drained water accumulated in villages in the southern part of the ring levee where the altitude was less than 1 m above sea level (hereinafter referred to as the lower villages) (Kaizu Town 1970a, p. 47). Debris from the upstream area was so great that the riverbed became higher than the surface of the land protected by the ring-levee, which meant it was very difficult to pump the drainage water out of the levee (Bekki 1932, pp. 248–249; Katano 1941, pp. 55–56). Therefore, the drainage water brought residents in the lower villages a higher risk of flooding and had negative effects on their rice farming. As a result, the residents of the Takasu ring-levee shared a common benefit in terms of flood control, but faced a conflict of interests in terms of artesian wells and drainage water (Nakazawa et al. 1936, p. 390).

The Kabu-ido system was developed to mitigate the conflict of interests associated with the drainage water. This system was essentially a permit system in which the total number of wells was restricted and residents who desired wells were required to pay a fee to obtain Kabu (the right) to drill a well. The revenue generated from these fees was then used to build and maintain drainage gates (Nakazawa et al. 1936, pp. 390–391; Matsubara 1968, p. 501; Ito and Aoki 1987, pp. 188–190).

8.2.2 The Kabu-ido System and the Coase Theorem

From a theoretical point of view, the Kabu-ido system was an attempt by local residents to mitigate negative effects related to drainage water through private negotiations. In this sense, the Kabu-ido system can be regarded as an example of the Coase theorem (Coase 1960). Specifically, the underlying cause of this problem was the lack of clear property rights regarding local groundwater. The poorly defined property rights resulted in an unspecified number of people being able to use the groundwater. Although drainage water produced by an individual groundwater user might be small, the accumulation caused a large problem. When an actor directly causes a negative impact on another actor without any compensation, these effects are known as "negative externalities," which can lead to inefficient resource allocation (Scitovsky 1954, pp. 145–146; Hardin 1968).

According to the Coase theorem, if certain conditions are met, private negotiations between parties can solve an externality problem without regard to the initial allocation of property rights. For example, if groundwater is the property of the lower villages, the upper villages can use groundwater so long as the lower villages allow it. However, if the upper villages take groundwater without permission and cause harm to the lower villages, they must compensate them. Such compensation will increase the cost of groundwater pumping, which should cause the volume of groundwater used to decrease. Conversely, if the groundwater is the property of the upper villages, the lower villages have to ask the upper villages to reduce groundwater pumping and compensate the upper villages by providing payments. The lower villages are willing to pay so long as the benefits caused by the reduction of groundwater pumping are greater than the payments. Under this scenario, groundwater pumping and the accompanying negative externality can be decreased, regardless of who owns the groundwater (Coase 1960).

In case of the Kabu-ido system, the upper villages were required to pay compensation to the lower villages. To determine why private negotiations worked and the upper villages agreed to pay the lower villages, it is necessary to consider how the Kabu-ido system developed.

8.3 Development of the Kabu-ido System

The Kabu-ido system was the outcome of private negotiations between the upper and lower villages. The development process can be classified into four stages: (1) monitoring, (2) adaptive management, (3) rule making, and (4) rule enforcement (Table 8.1).

8.3.1 Monitoring

The drought of 1852–1853 was the primary reason for the increase in artesian wells in the upper villages. Because of the adverse effects of drainage from the upper villages, the lower villages petitioned the local government for flood control and

Table 8.1 Development of the Kabu-ido system in the Takasu ring-levee

Year	Events
1852–1853	Drought hit the Takasu ring-levee
1854	A survey was conducted to determine the number of wells
1854	The upper and lower villages agreed on well restrictions and a 3-year adaptive management plan
1860	The upper and lower villages agreed on numbering of the wells and construction of a drainage gate
1861	The number of authorized wells was 388
1876	The maximum number of authorized wells increased to 806
1883	The total number of wells became 1,421
1905–	Kabu-ido gradually disappeared after construction of modern drainage gates

Source: Documents of Flood Control Office, Kasamatsu-jinya, Mino-gundai (1861), Katano (1941), Mori (1964), Matsubara (1968)

abolishment of wells in the upper villages (Documents of Flood Control Office, Kasamatsu-jinya, Mino-gundai 1854a, reprinted in Hirata Town 1984, p. 579). The local government had a policy of managing water problems outside a ring-levee (i.e., flood control) but not interfering with water issues inside a ring-levee (i.e., drainage problems) (Kaizu Town 1970a, p. 47). Although flood control and drainage problems were regarded as the most important water issues, conflicts inside a ring-levee that might negatively impact flood control that were a common problem to the upper and lower villages. As a result, when the local government implemented policies within ring-levees, they required unanimous consent from all villages (Matsubara 1968, p. 502). This policy led to the drainage problem being addressed via private negotiation among villages rather than government intervention.

However, the government did order the upper villages to check the number of artesian wells in response to a request from the lower villages. As a result, 529 wells (472 wells in operation and 57 abandoned wells) were found in a survey of June 1854. Figure 8.3 describes the location of wells at this time (Fig. 8.3) (Documents of Flood Control Office, Kasamatsu-jinya, Mino-gundai 1854b, reprinted in Mori 1964, pp. 920–924).

8.3.2 Negotiations Among Parties

In November 1854, well restrictions and a management plan were implemented based on agreements between the upper and lower villages. One of these agreements included prohibiting the use of irrigation wells during summer, because the discharge of artesian wells tended to increase during summer in response to increased water pressure from river flow (Matsuo 1993, p. 74). This agreement also reflected the traditional "try and see" method of resolving conflicts over water use. In the method called "Mi (see)-tameshi (try)" in old Japanese, problems are solved via a feedback process regarding implementation of various options and evaluation of the results. In modern terms, the method is referred to as adaptive management. In the agreement,

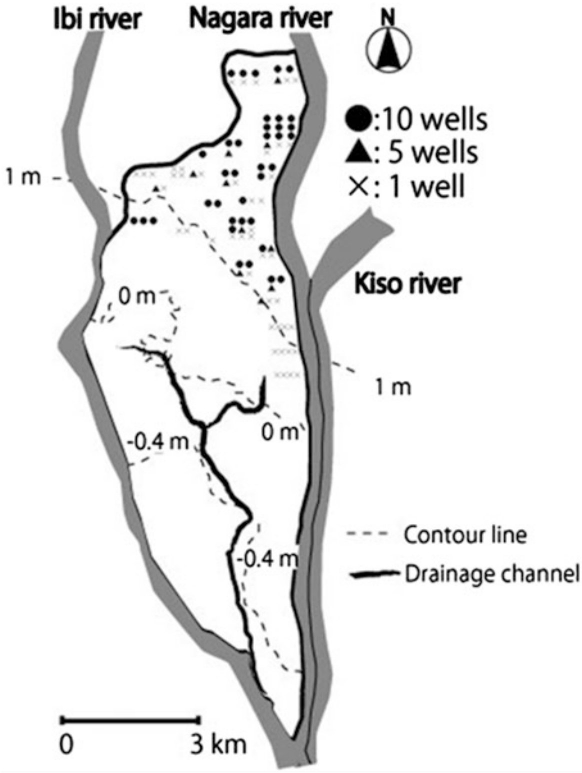

Fig. 8.3 Location of wells in operation as of June, 1854. [*Source*: Documents of Flood Control Office, Kasamatsu-jinya, Mino-gundai (1854b), reprinted in Mori (1964), pp 920–924]

both villages agreed on three points. First, they would check the magnitude of damage after 3 years and then determine whether the wells should be abolished. Second, representatives would be selected from both groups of villages for later negotiations. Third, negotiations would be conducted twice a year and the results would be submitted to the local government (Documents of Flood Control Office, Kasamatsu-jinya, Mino-gundai 1854c, reprinted in Mori 1964, p. 925; Documents of Flood Control Office, Kasamatsu-jinya, Mino-gundai 1855, reprinted in Mori 1964, p. 925.).

8.3.3 Rule Making

In 1860, the upper and lower villages started negotiations again after a severe inundation of the lower villages following a series of heavy rainfall events. As a result, the following agreement was made between the upper and lower villages: (1) building drainage gates at the expense of the upper villages (Documents of Flood Control Office, Kasamatsu-jinya, Mino-gundai 1860a, reprinted in Kaizu Town 1970b, p. 780), and (2) numbering of the wells in the upper villages (Documents of Flood Control Office, Kasamatsu-jinya, Mino-gundai 1860b, reprinted in Hirata Town 1984, p. 597). The numbering led to easy visualization of the total number of wells and made it difficult for the upper villages to dig unpermitted wells because the lower villages could easily detect hidden wells by checking wells without a number or with a duplicate number. As Mori pointed out (Mori 1964, p. 934), the original form of the Kabu-ido system could be seen at this stage. According to a survey done in 1861, the total number of wells were 542 (388 wells in operation and 154 abandoned wells). The 388 wells that were in operation at the time were all numbered. Moreover, the number of wells was severely restricted as compared to the one in operation in June 1854 (i.e., 472). Therefore, it can be inferred that the Kabu-ido system was launched at this stage (Documents of Flood Control Office, Kasamatsu-jinya, Mino-gundai 1861).

8.3.4 Rule Enforcement

The agreement of 1860 included the following articles of enforcement:

- No additional wells could be dug in the upper villages without permission from the lower villages.
- If the upper villages wanted to open a well under seasonal restriction, they were required to go to the local government with the residents of the lower villages and ask for their advice. This rule is enforced even during droughts.
- If a resident of the upper villages opens a well under seasonal restriction without permission, his/her well must be destroyed and they were fined 5 Ryo. If he/she cannot afford to pay, the village must pay on their behalf. [Ryo was a unit of money at that time; 1 Ryo in 1860 was equivalent to 4,000 Japanese yen or US $50 (Ikeda and Hayashiya 1964, p. 204).]
- 2 Ryo of 5 Ryo can be given as a reward to a person who finds an unpermitted well. The remainder (i.e., 3 Ryo) is used for maintenance of the drainage gates in the lower villages. (Documents of Flood Control Office, Kasamatsu-jinya, Mino-gundai 1860a, reprinted in Kaizu Town 1970b, p. 783; Documents of Flood Control Office, Kasamatsu-jinya, Mino-gundai 1860b, reprinted in Hirata Town 1984, p. 597.)

The combination of a penalty system with a reward system demonstrates how severe was the conflict between the upper and lower villages. Natural resources are often used excessively, even when there are rules regarding their utilization. The Kabu-ido system reflects the importance of enforcing the rules. Although it was not perfect, the enforcement aspect of the Kabu-ido system worked to some degree because some hidden wells were actually destroyed (Mori 1964, pp. 935–936).

After 1903, the Kabu-ido system gradually disappeared as modern drainage pumping machines prevailed in the Takasu ring-levee. It enabled the lower villages to pump out excess drainage water easily, and restriction of groundwater pumping was no longer necessary. This difference shows that the Kabu-ido system was developed under technological constraints (Matsubara 1968, p. 501).

8.4 Discussion: Implications for Current Groundwater Management Policy

As mentioned in the previous chapter, the groundwater problem in the Takasu ring-levee was solved through (1) monitoring, (2) negotiations among parties, (3) rule making, and (4) rule enforcement. This process can be applied to various groundwater problems that currently exist. Moreover, some policy implications for modern groundwater management can be deduced from the Kabu-ido system.

8.4.1 Importance of Third-Party Monitoring

As previously mentioned, the local government ordered the upper villages to check the number of artesian wells in response to requests from the lower villages. According to a report submitted by the upper villages to the government in March 1854, the total number of wells was 352 (including six abandoned wells) (Documents of Flood Control Office, Kasamatsu-jinya, Mino-gundai 1854d, reprinted in Hirata Town 1984, pp. 577–578). However, the local government found 529 wells (including 54 abandoned wells) in its own survey conducted half a year later (Documents of Flood Control Office, Kasamatsu-jinya, Mino-gundai 1854b, reprinted in Mori 1964, pp. 920–924). According to Matsubara, this difference was likely because the former survey was conducted by the upper villages and the latter by the local government. The upper villages had an incentive to report fewer wells than they actually observed (Matsubara 1968, p. 495). This point of view can be supported by comparison of two reports. The report by the upper village merely determined the total number of wells based on the distinction between old and new wells. Conversely, the latter report attempted to grasp the current situation by including the purposes of wells (irrigation or drinking water) and counting abandoned wells precisely.

This interpretation suggests that monitoring the current situation should be conducted by a third party that is not a direct stakeholder. Indeed, monitoring by a third party will reduce transaction costs and promote private negotiations.

8.4.2 Issue Linkage

According to the Coase theorem, private negotiation between the upper and lower villages can take either of two forms: the upper villages pay the lower ones compensation for the damage caused by their wells or the lower villages ask the upper village to reduce groundwater pumping in return for payment. Negotiations resulted in the former case in the Kabu-ido system described herein. Although the reason for this is not documented, it is likely that the need for flood control led to this outcome. Although the upper and lower villages were in conflict over groundwater pumping inside the ring-levee, they had a common need to protect themselves against occasional flooding. The upper villages could not implement effective protection against floods without the help of the lower villages. Additionally, damage caused by the collapse of a levee can be very large and widespread. Accordingly, it is likely that the residents of the upper villages considered the benefit of protecting their lives and property to be greater than the compensation they paid to the lower villages.

This observation suggests the importance of "issue linkage" or "issue packaging." The more issues that parties discuss, the easier it is for them to make compromises, because even if a party loses benefits on one issue they are likely to recover the loss on another issue. If restriction of groundwater pumping were the only issue, the upper village would have no incentive to accept well restrictions. However, if issues regarding well restriction and flood control were discussed at the same time, the upper villages might have greater incentive to compromise on well restrictions. Accordingly, it can be inferred that such "issue packaging" promoted consensus building between the upper and lower villages.

8.4.3 Wastewater Charge

The Kabu-ido system promoted internalization of external diseconomy related to drainage water by well restriction and compensation. The Kabu-ido system can also be considered a system of wastewater fee. The charge increased the cost of well construction and provided the upper villages with an incentive to reduce groundwater pumping.

In addition to charging for wastewater, application of a pump tax can also be used to restrict groundwater abstraction, and both methods can play a role in internalization of such external diseconomies. The pump tax method promotes internalization at the beginning of groundwater abstraction. A good example of this is the groundwater management policy in Bangkok, Thailand, which was introduced in 1977 to restrict land subsidence (Endo 2011). Conversely, the land subsidence policy in Hiratsuka, Kanagawa Prefecture, Japan is a good example of application of a wastewater charge system. As Shibazaki showed, introduction of a wastewater charge system played a role in stopping land subsidence because it induced the industrial sector, which was the main user of groundwater at the time, to use recycled water to avoid the charge (Shibazaki 1981, pp. 61–63).

Both the pump tax and wastewater charge system can be useful in restricting groundwater pumping; however, there is a large difference between these measures with respect to policy making. Specifically, although the former is subject to a rule of land ownership, the latter is not. In some countries, groundwater is considered to be part of the land it underlies. In such cases, it is politically and legally difficult for policy makers to introduce a pump tax because it will be criticized as infringement of land ownership. This problem can be avoided in

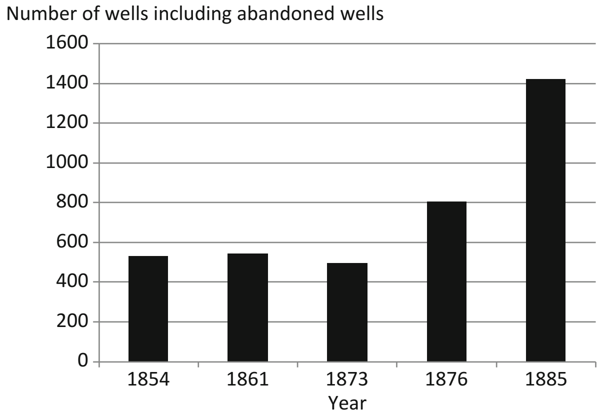

Fig. 8.4 Change of well number in Takasu ring-levee. [*Source*: Documents of Flood Control Office, Kasamatsu-jinya, Mino-gundai (1861); Mori 1964; Matsubara (1968)]

wastewater charge systems. In this sense, the consideration of the Kabu-ido system leads to enlargement of the policy options available for groundwater management.

8.4.4 Establishing a Fund with a Specific Purpose

As previously mentioned, 388 wells were numbered in 1861, but 406 wells were found in a survey of October 1873 (Documents of Flood Control Office, Kasamatsu-jinya, Mino-gundai 1873). Therefore, about 20 unauthorized wells were made during this period. The lower villages asked the upper ones to abolish those wells and the latter accepted the request (Mori 1964, pp. 937–938; Matsubara 1968, p. 498).

However, drought occurred again and new wells were added in the upper villages without permission from the lower villages. To address this problem, both villages reached the following compromise: the lower villages allowed the total number of wells in the upper villages to increase to 806 and the upper villages paid for construction of three new drainage gates. In addition, sanctions against rule breaking were discussed and both sides agreed that if the upper villages constructed additional wells without permission from the lower villages all existing wells would be destroyed (Mori 1964, pp. 940–942). In 1885, 806 wells were decreased to 421 and an additional 1,000 wells were proposed in exchange for improving the drainage gates in the lower villages in the same year. As a result, the total number of the wells became 1,421 (Mori 1964, pp. 942–945; Matsubara 1968, pp. 499–500) (Fig. 8.4).

As this agreement shows, the main focus of the Kabu-ido system moved from limitation of the total number of wells to fund raising through the sale of well permits, suggesting that imposing pumping taxes and wastewater charges does not always have a part in restricting groundwater abstraction. Indeed, such policies may be used only to secure a regional budget. In the case of the Kabu-ido system, funds raised in exchange for well permits were used for the drainage gates in the lower villages, resulting in mitigation of the adverse effects of artesian wells. As no governmental subsidies were available for construction and maintenance of the drainage gates, the residents had no choice but to manage the drainage gates at their own expense (Matsubara 1968, pp. 499–501). This point suggests that economic tools such as pump taxes and wastewater charges should be arranged in the form of an ear-marked tax to ensure consistency of the policy.

8.5 Conclusion

The purpose of this chapter was to reconsider the Kabu-ido system that once existed in the southern part of the Noubi Plain, Tokai, Japan, and to deduce policy implications for current groundwater management policy. Kabu-ido was a customary groundwater management system designed to mitigate drainage problems caused by excessive groundwater pumping. The system had two features: limitation of the total number of wells and groundwater users' obligation to pay economic compensation.

The Kabu-ido system is as an example of the Coase theorem, because it can be regarded as an institution by which the stakeholders internalized negative externalities caused by groundwater pumping through private negotiations.

Although the Kabu-ido system in the Takasu ring-levee emerged in the 1860s and disappeared around 1905, it still contains implications for current groundwater management policy including the importance of third-party monitoring, usefulness of issue linkage in the process of conflict management, effectiveness of economic tools on the use of groundwater pumping restrictions, and significance of an ear-marked tax.

A major cause of excessive groundwater pumping is poorly defined property rights for groundwater resources. However, it is difficult to set up and enforce clear property rights for groundwater because the exact extent is not always clear. Residents within the Takasu ring-levee faced the problem of determining how to restrict the number of groundwater users, pumping volume, and well construction technology through institutional arrangements. These problems are similar to current problems such as land subsidence and groundwater pollution. Accordingly, the Kabu-ido system, which was used more than 100 years ago, still provides useful information that is applicable to current groundwater management policy.

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All the historical documents on the Kabu-ido system cited in this chapter (Documents of Flood Control Office, Kasamatsu-jinya, Mino-gundai) are available in the Gifu Prefecture historical document center. The views presented here are those of the author and should not be attributed to Gifu Prefecture in any way. Responsibility for the text (with any surviving errors) rests entirely with the author.

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Documents of Flood Control Office, Kasamatsu-jinya, Mino-gundai (1861) Survey of working wells and abandoned wells

Documents of Flood Control Office, Kasamatsu-jinya, Mino-gundai (1873) Survey of artesian wells in Takasu ring-levee in 1873

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