Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

At the end of this chapter, students expect to know:

- **Latitudinal upper atmosphere temperature and wind variation**
- **Types of balanced wind flow**
- **Tropopause height variation**
- **Type of Jet stream and its characteristics**
- **Cyclone and anti cyclone pressure system**

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

Tropospheric and Stratospheric temperature

Troposphere is the lowest part of the atmosphere and is closer to the Earth, and extends about 8 km above the poles and 18 km over the equator. Troposphere means region of turning or mixing

The densest part of the atmosphere in terms of water vapor, clouds, and precipitation,airmass,concentration.

Temperature generally decreases with height in the troposphere at about 6–7◦Ckm−1 inthelowerhalfand7–8◦Ckm−1 in the upper half. Because of the general decrease of temperature with height and the presence of weather systems, the troposphere is often characterized by fairly significant localized vertical motions.

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

The temporal and latitudinal variability of mean temperature profile shows that there is considerable latitudinal and seasonal variability.

Temperature decreases with latitude in the troposphere.

The latitudinal gradient is about twice as steep in the winter hemisphere compared to that in the summer hemisphere.

The latitudinal distribution of temperature in the lower Stratosphere is rather complicated.

The summer hemisphere has a cold equator and a warm pole.

The winter hemisphere is cold at both equator and pole with a warmer region in middle latitudes.

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

The cold pool of stratospheric air over the winter pole is highly variable.

At the stratopause, there is a monotonic temperature gradient between the warm summer pole and the cold winter pole.

During these so-called sudden stratospheric warmings, the stratospheric temperatures over individual stations have been observed to rise by as much as 70◦C in 1 week

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

Water vapor plays a major role in regulating air temperature because it absorbs solar energy and thermal radiation from the planet’s surface. The troposphere contains 99% of the water vapor in the atmosphere.

Water vapor content, decreases rapidly with altitude.

The troposphere is bounded at the top by the tropopause, whose altitude varies considerably depending on the location and type of weather systems, latitude

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

Tropopause is the boundary b/n troposphere and stratosphere

It varies with weather system.

The tropopause is an important meteorological concept.It separates the troposphere from the stratosphere,i.e., two volumes of air with significantly different properties.

In this region the air ceases to cool,and the air be comes almost completely dry. Basically, it is the boundary between the upper troposphere and the lower stratosphere that varies in altitude between the poles and the equator.

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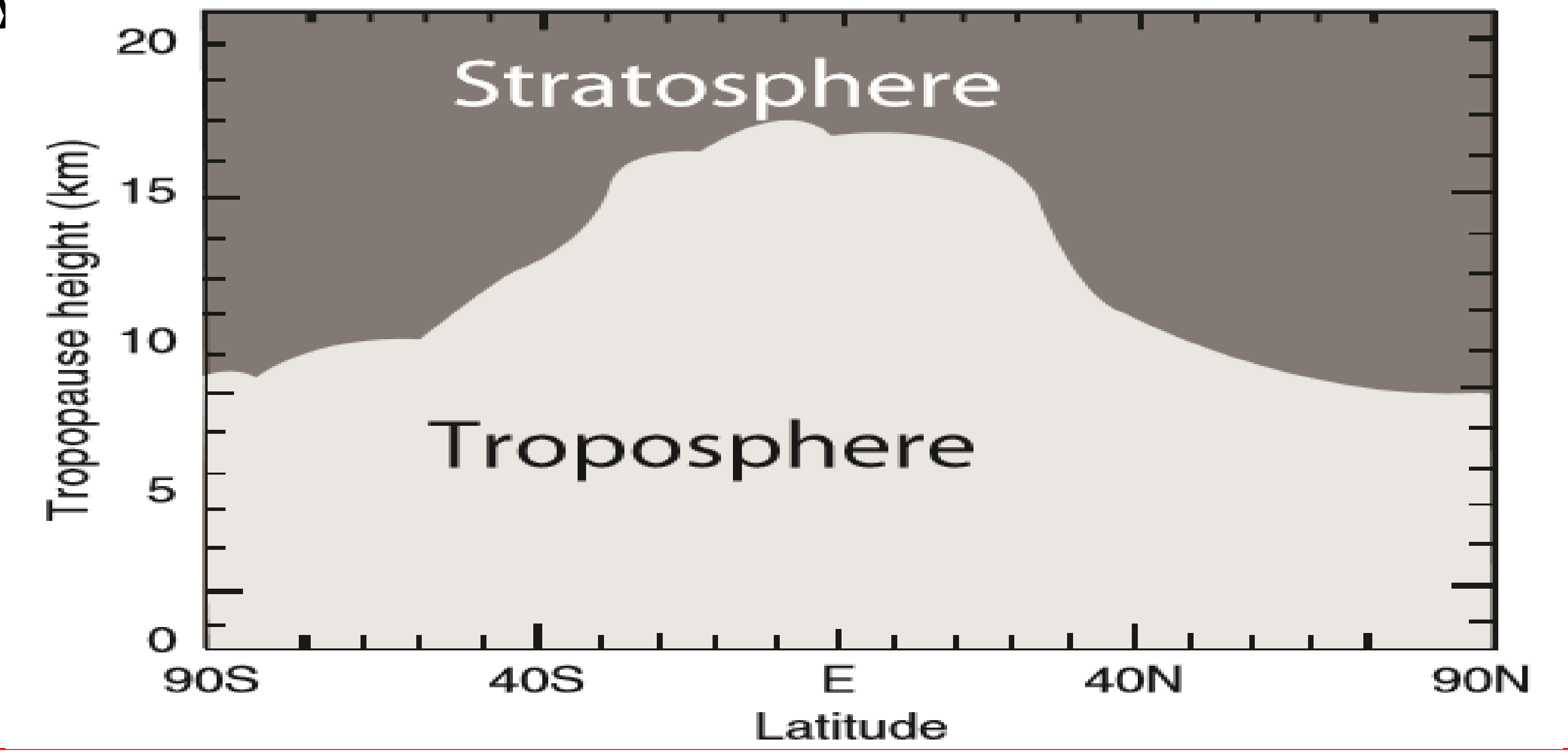
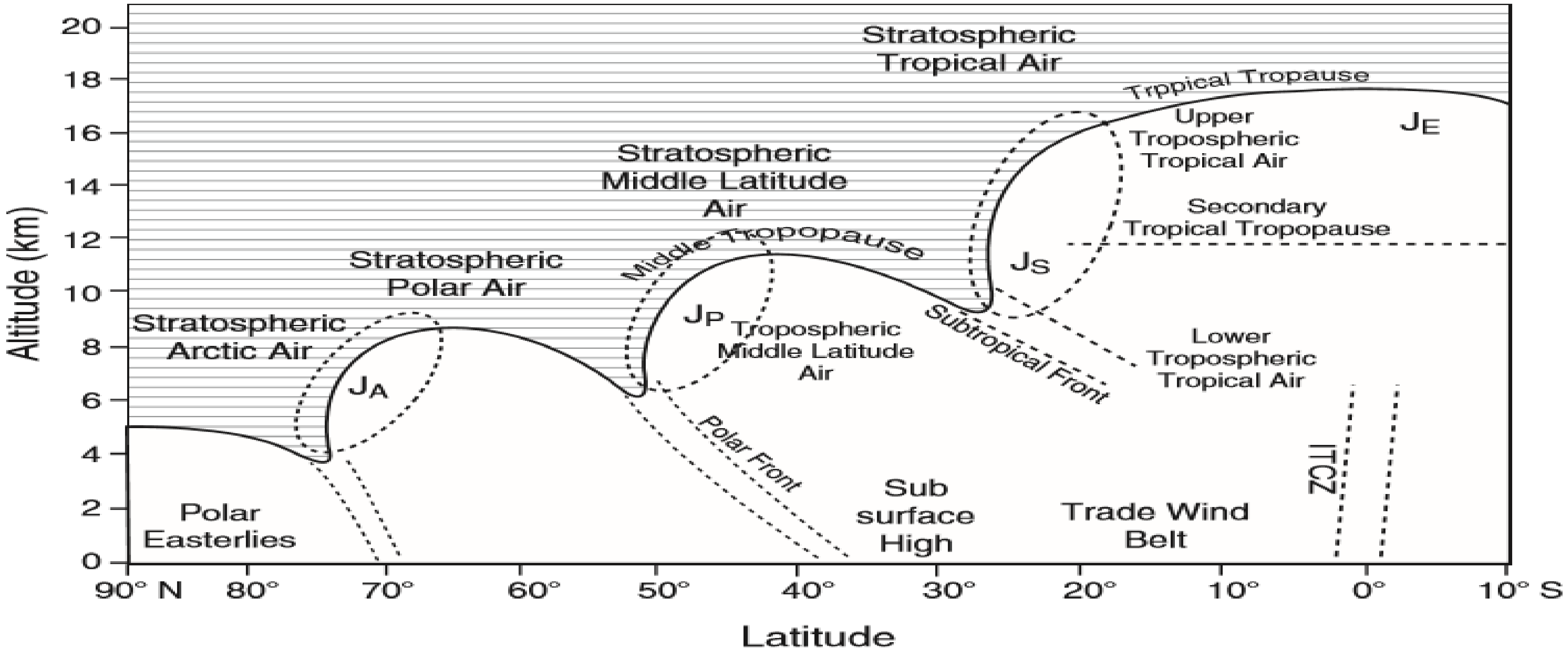
The tropopause is not a fixed boundary

Tropopause height shows large variations with latitude, season, and even day to-day



Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

The tropopause height also varies from troughs to ridges, with low tropopause height in cold troughs and high in warm ridges.



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Characteristic features of tropopause at various latitude zones

Features	Tropical tropopause	Midlatitude tropopause	Polar tropopause
Location	Over tropics, between the two subtropical jet streams	Between polar and subtropical jet streams	North of polar jet
Height	~18 km	~12 km	6–9 km
Altitude	~80–100 hPa	~200 hPa	~300–400 hPa
Temperature	~ −80°C	~ −60°C	~ −45°C
Potential temperature	~375–400 K	~325–340 K	~300–310 K
Character	Sharply defined, highest and coldest	Higher in summer and lower in winter	Often difficult to identify

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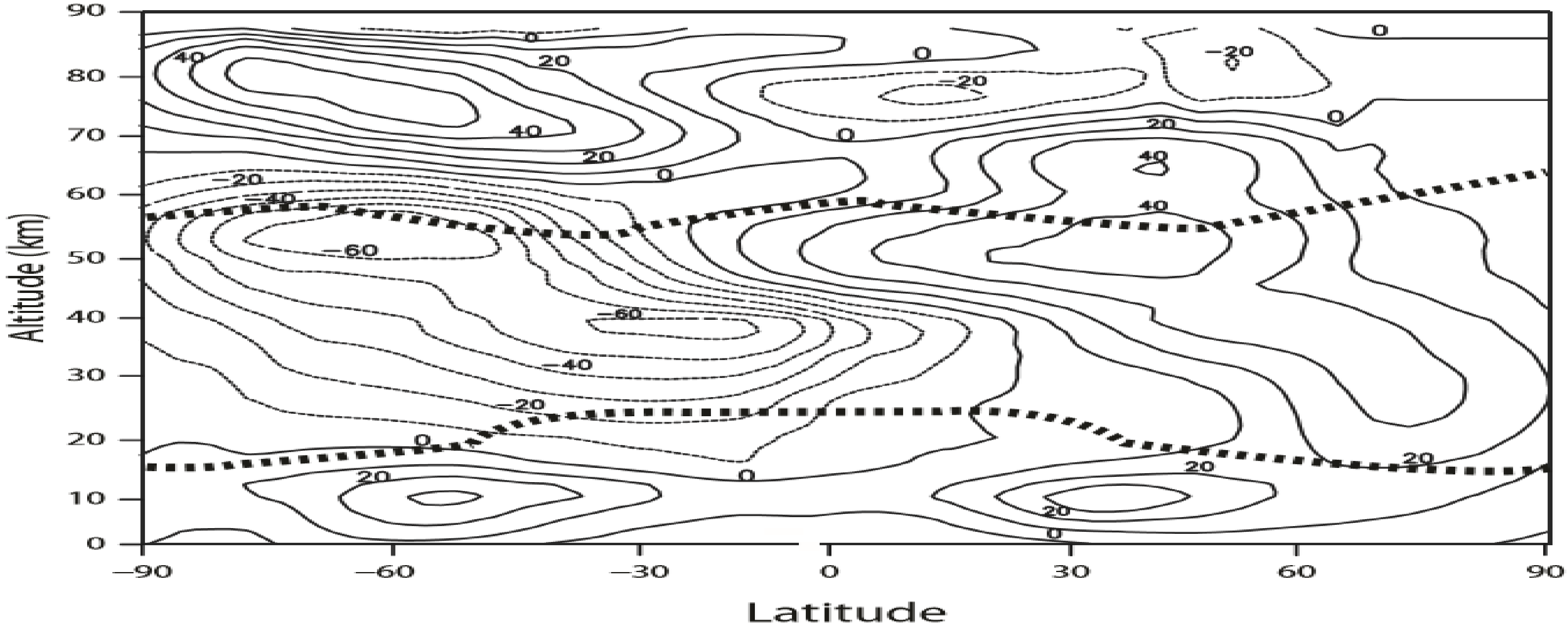
Tropospheric and Stratospheric wind

The large-scale features of the atmospheric zonal wind circulation from surface to 90km are shown in Figure.



Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

The mean zonal flow in the winter hemisphere equatorward of 40◦ latitude is similar, with stronger westerlies about40ms^{-1} at the200hPa level.

The maximum wind in the southern hemisphere (SH) is about 2–3◦ latitude nearer the equator and isabout5ms−1 weaker than the northern hemisphere (NH)winter maximum.

Poleward of 40◦S latitude, the zonal winds differ appreciably in winter, with stronger winds in the SH. A westerly maximum in the upper troposphere that continues into the stratosphere is evident between 50◦ and 60◦S in accordance with the upward increasing meridional temperature contrast poleward of 45◦S.

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

The distribution of wind differs considerably between the summer hemispheres.

The upper troposphere westerly maximum is nearly twice as strong in the SH and is located farther poleward than the peak in the NH.

In the middle and upper troposphere the tropical easterlies are much stronger in the NH than in the SH, and in the subtropics the westerlies are much stronger in the SH.

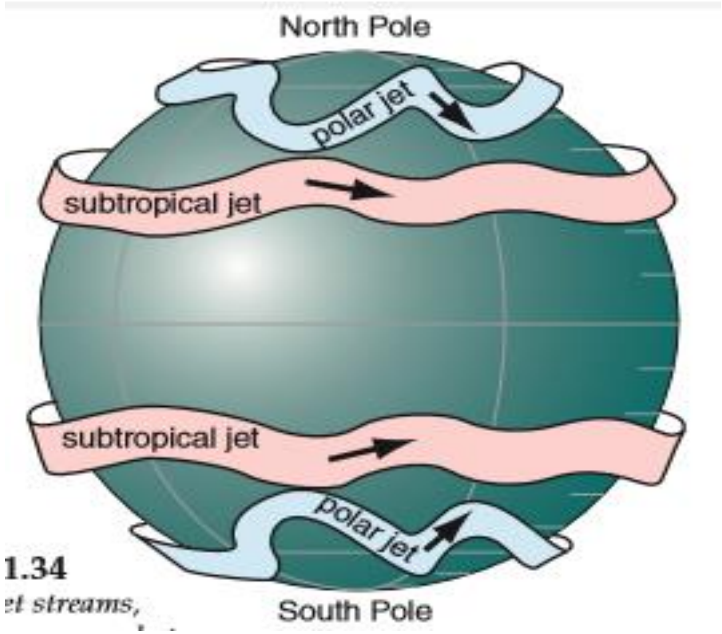
Prominent features are cores of strong westerly winds in middle latitudes at an altitude of 10 km. However, the strongest zonal winds occur in the mesosphere at an elevationof 60km.

Again there are two jet cores in middle latitudes, the stronger in the winter hemispheres westerly; the other in the summer hemisphere is easterly.

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

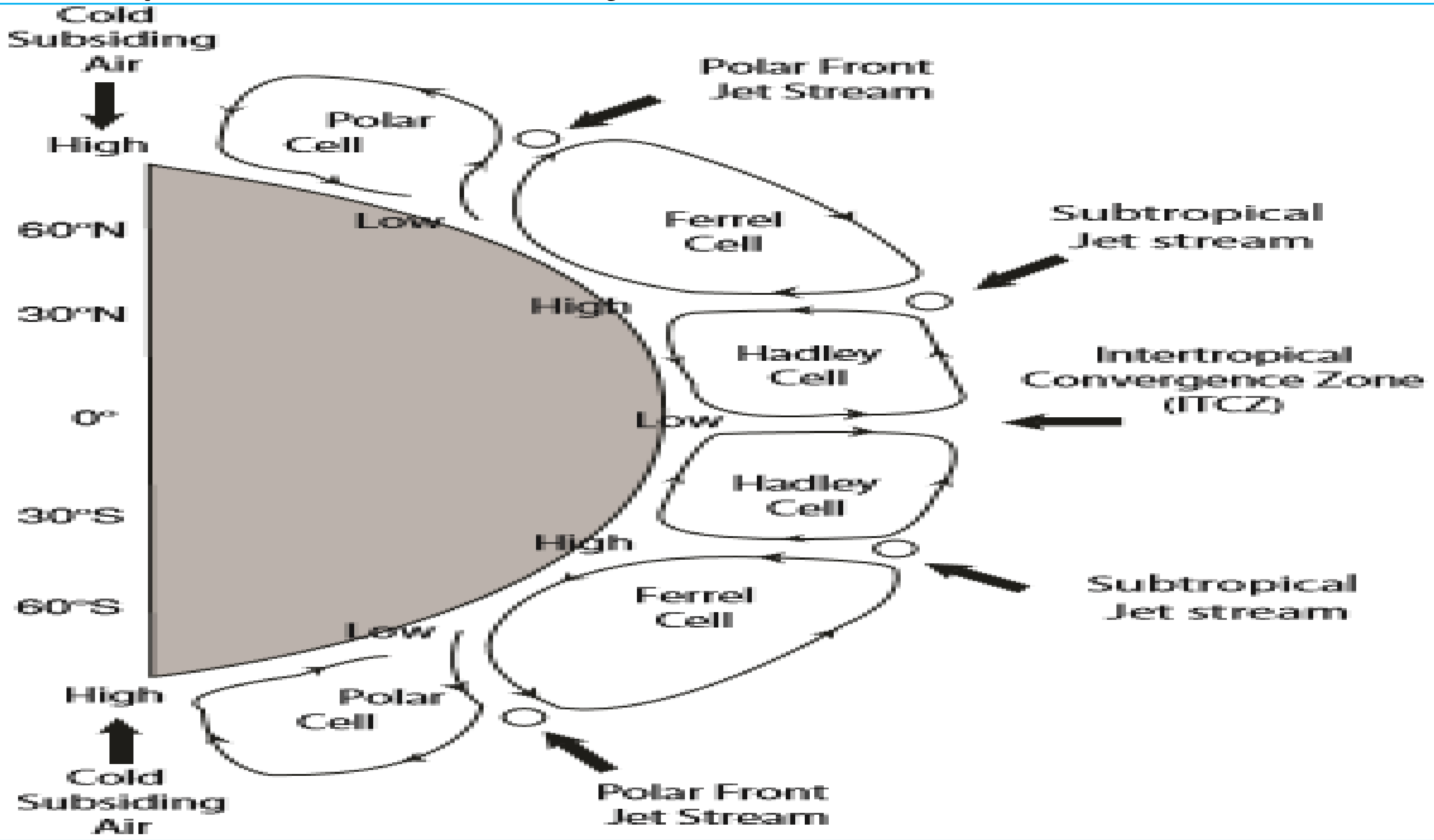
4.3Principal Tropospheric Jet Streams

In the winter hemisphere there are often two strong jet streams of fast west-to-east moving air near the tropopause: the polar jet stream and the subtropical jet stream.



Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

The three cells are responsible to form jet streams



Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

Characteristics of Jet stream

(1) is very steady;

(2) meanders north and south a bit;

(3) is about 10° latitude wide (width ≈1,000 km); and

(4) has seasonal-average speeds of about 45 m s^{-1} over the Atlantic Ocean, 55 to 65 m s^{-1} over Africa and the Indian Ocean, and 60 to 80 m s^{-1} over the western Pacific Ocean

The core of fast winds near its center is at 12 km altitude

It is driven by outflow from the top of the Hadley cell, and is affected by both Coriolis force and angular-momentum conservation.

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

The polar jet is centered near 50 to 60° latitude in the winter hemisphere.

Characteristics of polar jet:

(1) is extremely variable;

(2) meanders extensively north and south;

(3) is about 5° latitude wide; and

(4) has widely varying speeds (25 to 100 m s^{-1}) driven by varying horizontal temperature gradients.

The core altitude is about 9 km.

The transient meanders of the polar jet (troughs and ridges in the Rossby waves) are extremely important for mid-latitude cyclone formation and evolution.

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

In the summer hemisphere, the jets from the west have merged and the winds are slower because of the weaker temperature contrast between the equator and the warm pole.

Core wind speeds in the jet are 0 to 10 m s^{-1} in N. Hemisphere summer, and 5 to 45 m s^{-1} in S. Hemisphere summer.

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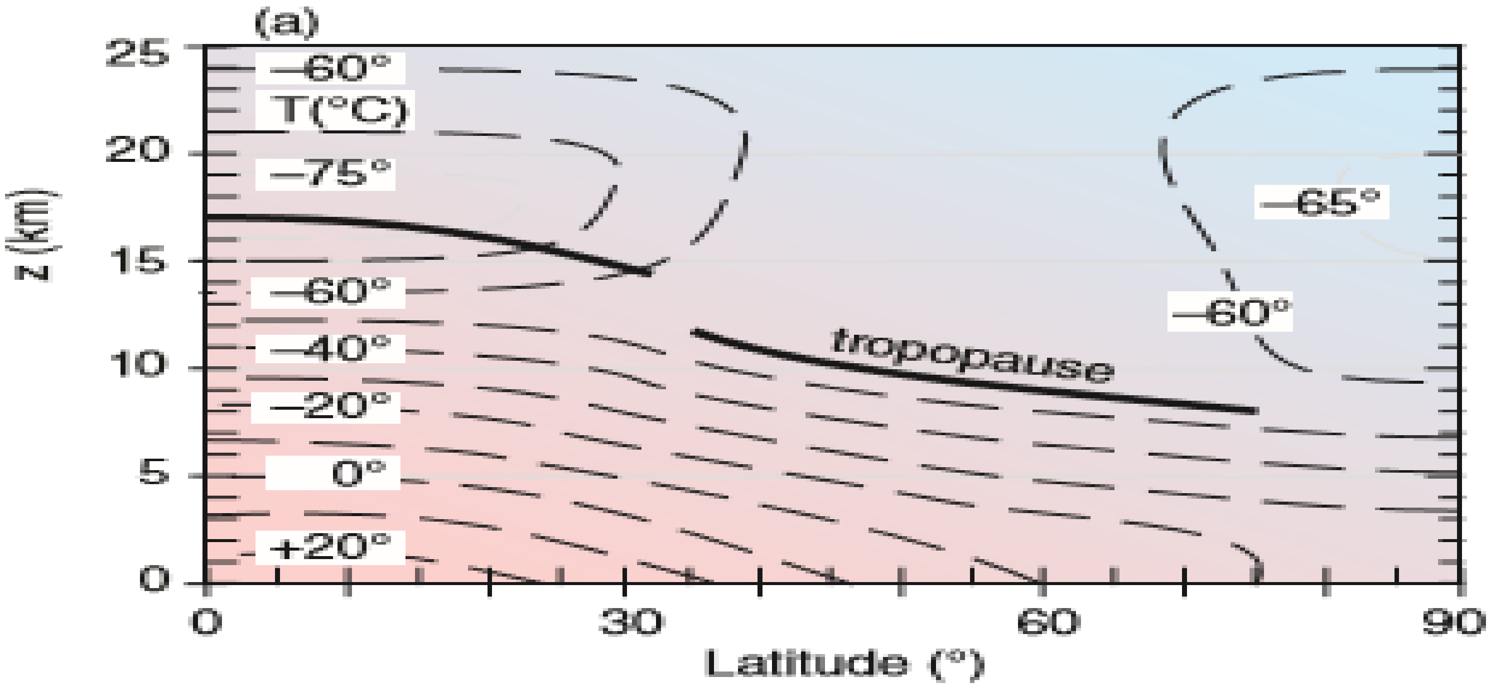
Baroclinicity & the Polar Jet:

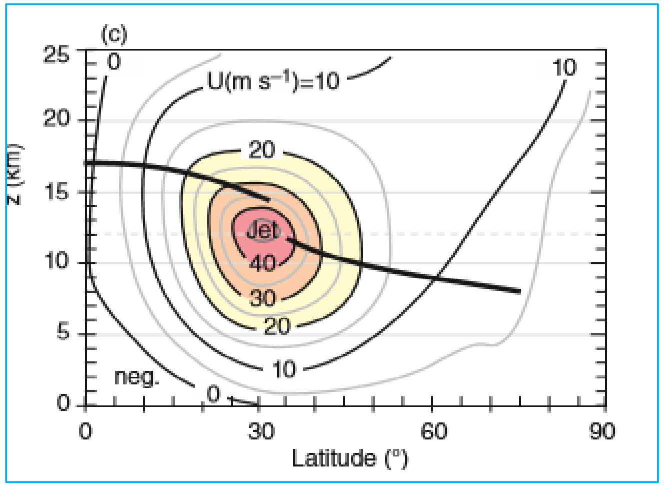
At any altitude in the troposphere you will find a horizontal temperature gradient between colder poles and warmer equator.

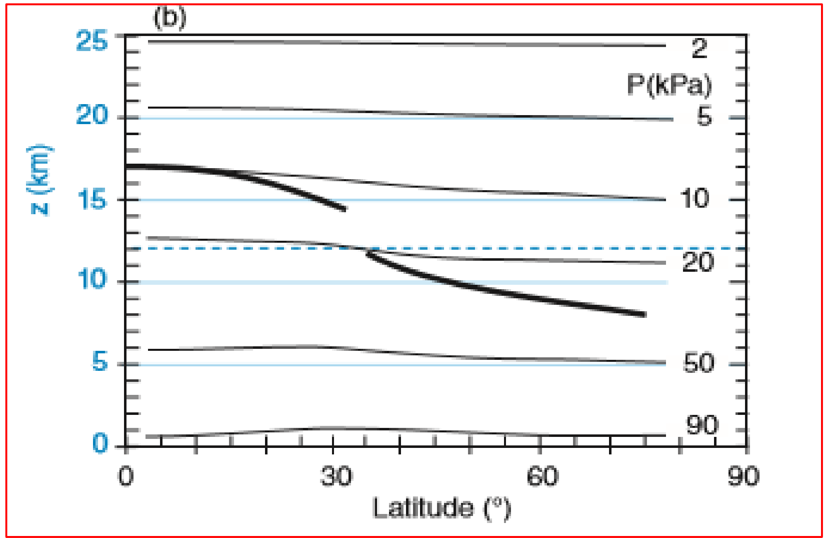


Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

According to the hypsometric relationship, the thickness between two isobaric surfaces is smaller in the colder (polar) air and greater in the warmer (equatorial) air.

Hence, isobaric surfaces tilt in the horizontal, which drives a geostrophic wind





Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

The greatest tilt is near 30° latitude at the tropopause, and the associated pressure gradient drives the fastest winds (jet-stream core) there, as expected due to the thermal wind.

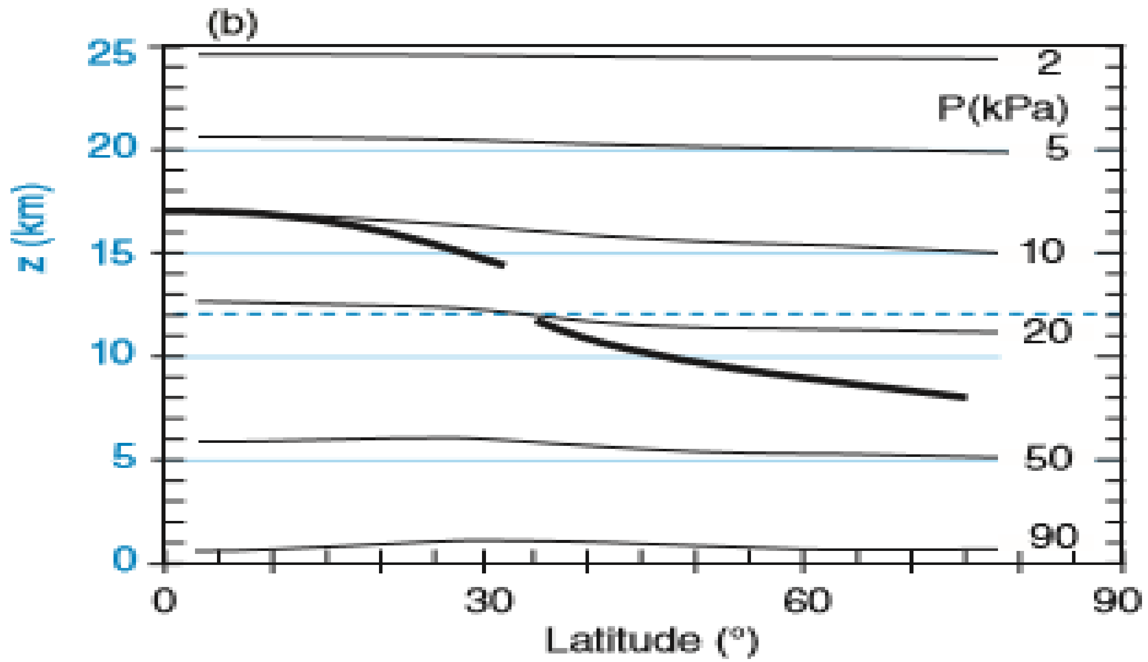
Because the isobars cross the isotherms (and isobars also cross the isopycnics — lines of equal density), the atmosphere is said to be baroclinic.

It is this baroclinicity associated with the meridional temperature gradient that creates the west winds of the jet stream

the troposphere is deeper near the equator than near the poles. Thus, the typical lapse rate in the troposphere, applied over the greater depth, causes colder temperatures at the tropopause over the equator than over the poles

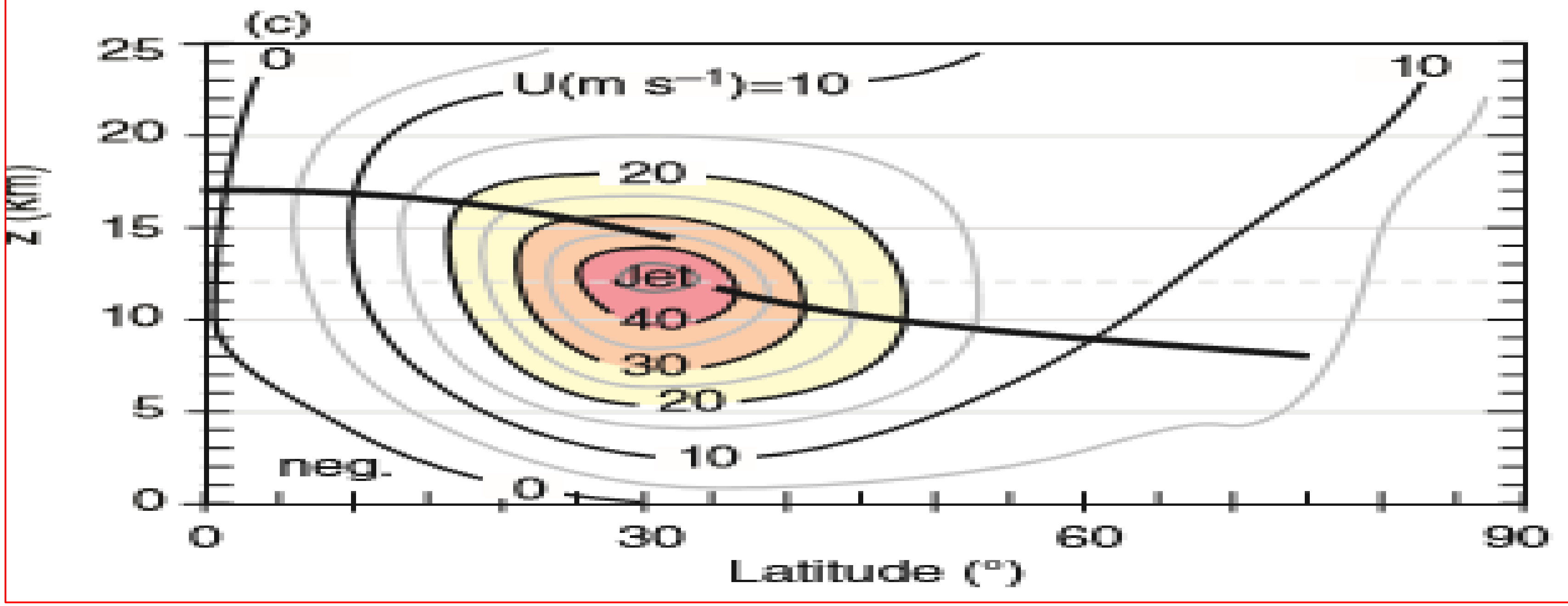
Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

The associated north-south thickness changes between isobaric surfaces cause the meridional pressure gradient to decrease as reductions in slopes of the isobars.



Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

The reduced pressure gradient in the lower stratosphere causes wind speeds to decrease with increasing altitude leaving the jet max at the tropopause.



Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

Angular Momentum & Subtropical Jet:

Angular momentum can influence the subtropical jet, and is defined as mass times velocity times radius of curvature.

Suppose that initially there is air moving at some zonal velocity U_s relative to the Earth's surface at some initial (source) latitude φ_s.

Because the Earth is rotating, the Earth's surface at the source latitude is moving toward the east at velocity U_{Es}.

Thus, the total eastward speed of the air parcel relative to the Earth's axis is $(U_s + U_{Es})$.

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

suppose some disturbance such as a meandering jet stream moves the air to some other (destination) latitude φd, assuming that no other forces are applied.

Conservation of angular momentum requires:

$$m \cdot (U_s + U_{Es}) \cdot R_s = m \cdot (U_d + U_{Ed}) \cdot R_d$$

where U_d represents the new zonal air velocity relative to the Earth's surface at the destination latitude, U_{Ed} is the tangential velocity of the Earth's surface at the destination, and m is air mass.

For latitude φ at either the source or destination, the radius is R_s or R_d = RE·cos(φ), where average Earth radius is RE = 6371 km.

Similarly, tangential velocities at either the source or destination are U_s or U_d = Ω·Rφ = Ω·RE·cos(φ), for an Earth angular velocity of $\Omega = 0.729 \times 10^{-4}\ s^{-1}$.

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

Near 30° latitude in each hemisphere is a persistent belt of strong westerly winds at the tropopause called the subtropical jet.

In mid-latitudes at the tropopause is another belt of strong westerly winds called the polar jet. The centerline of the polar jet meanders north and south, resulting in a wave-like shape called a Rossby wave.

Jet streams are fast flowing air currents, which are 1000's of km long, few 100 km wide, one a few kilometers thick.

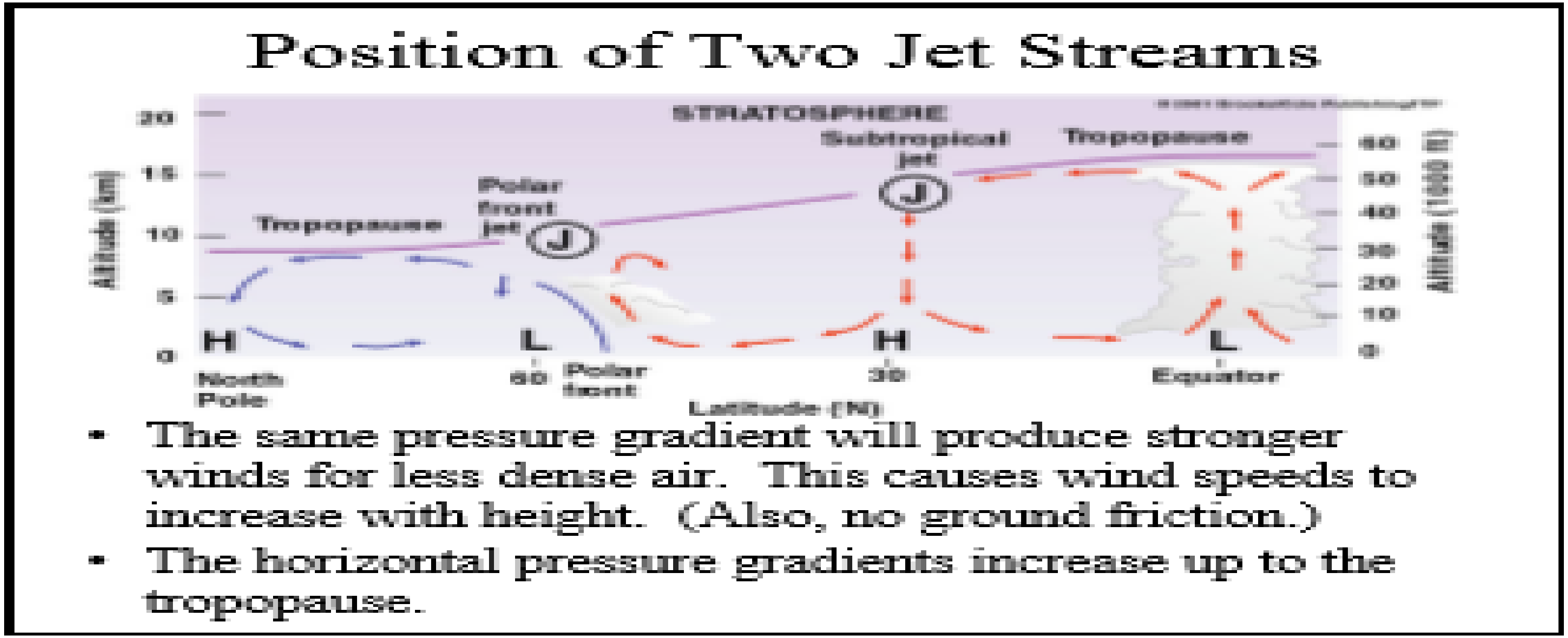
Wind speeds at the core often exceed 100 knots, and occasioanlly exceed 200 knots.

They are usually found in the tropopause between 10 and 15 km in elevation.

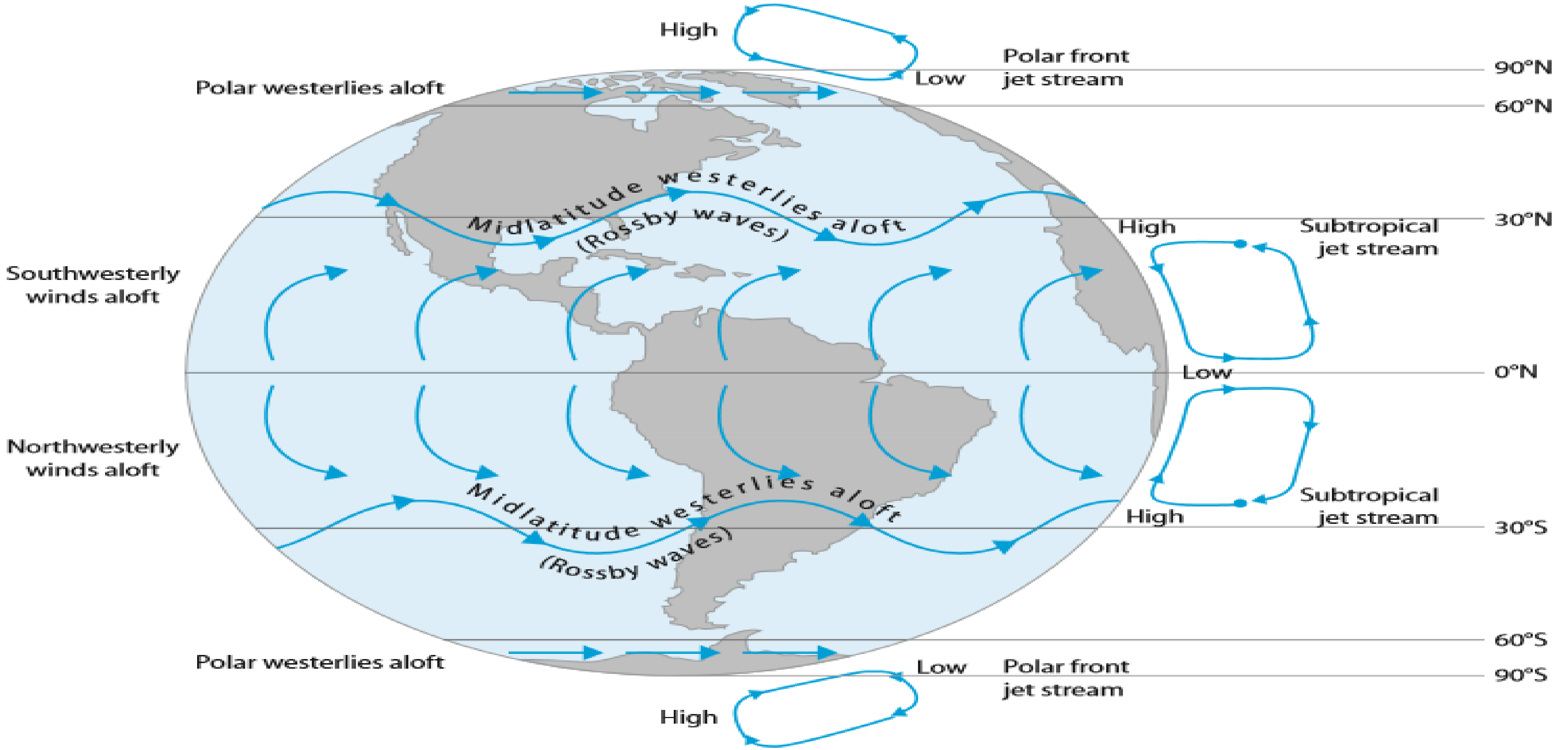
Weaken during summer and strengthen in winter

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

Position of Two Jet Streams



Chapter 4 General Circulation Upper Atmospheric and Flow Patterns



Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

FORMATION OF UPPER-TROPOSPHERIC CYCLONES AND ANTICYCLONES

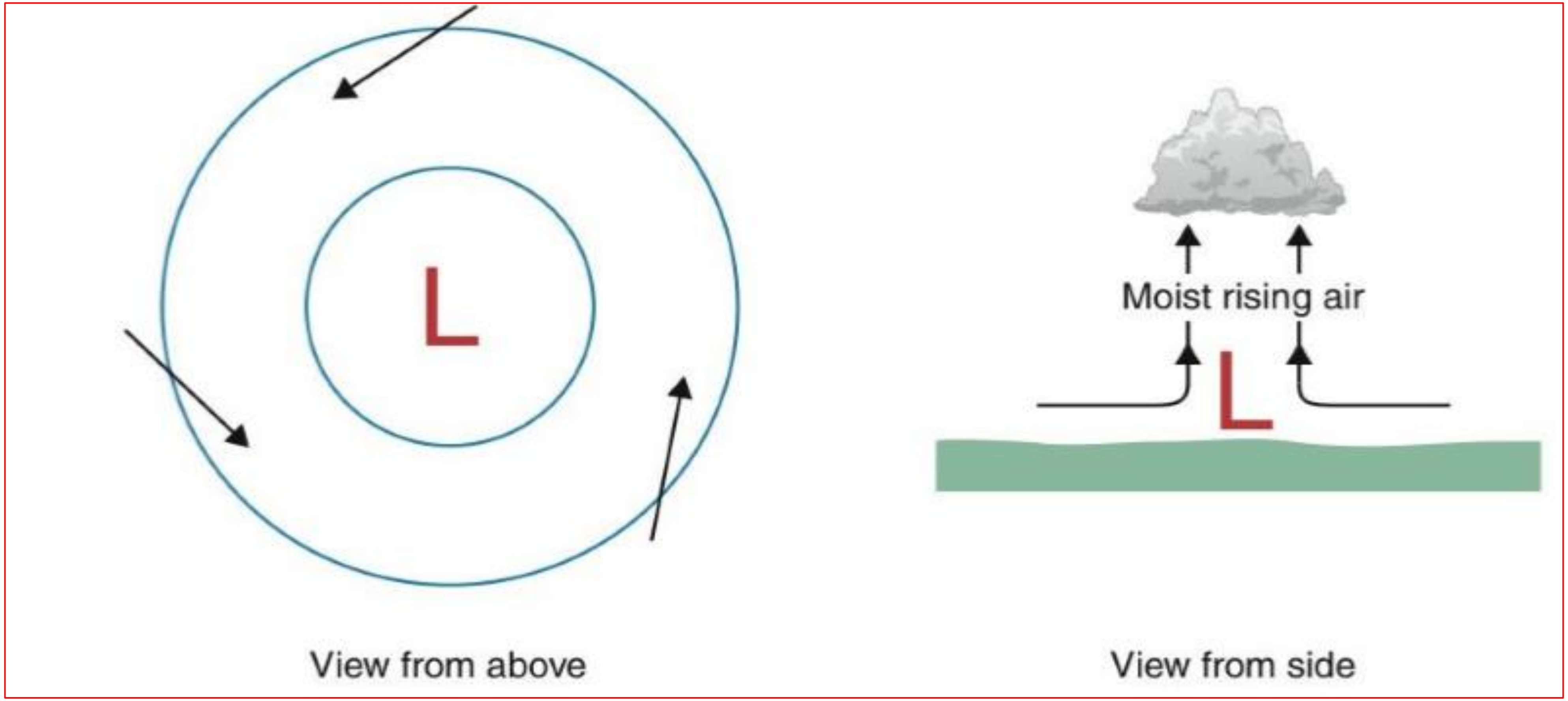
A cyclone is simply an area of low pressure around which the winds flow counterclockwise in the Northern Hemisphere and clockwise in the Southern Hemisphere.

- Cyclones form and grow near the front
- Cyclones (lows) are cloudy, wet, stormy
- Cyclones have converging air at surface that rises

The deformation of the middle- and upper-tropospheric flow connected with the deepening of waves in the westerlies often results in the formation of closed lows equatorward (and closed highs poleward) of the main belt of zonally averaged westerlies.

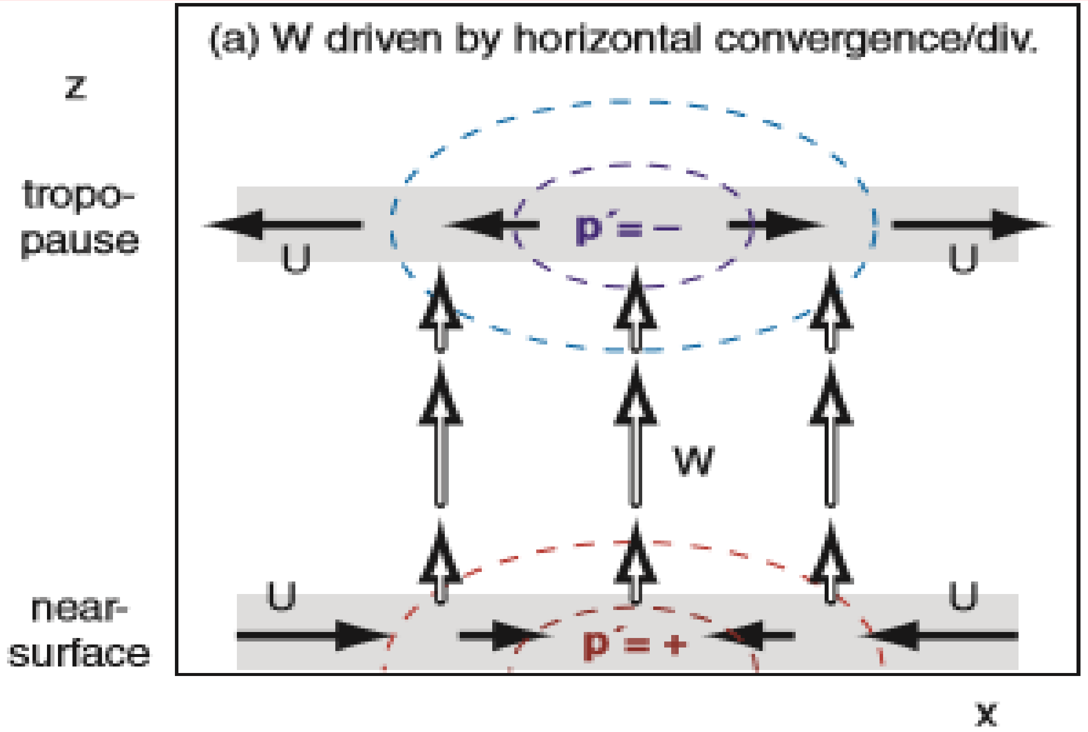
These lows and highs may then persist for quite a long time and have a profound influence on the weather.

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns



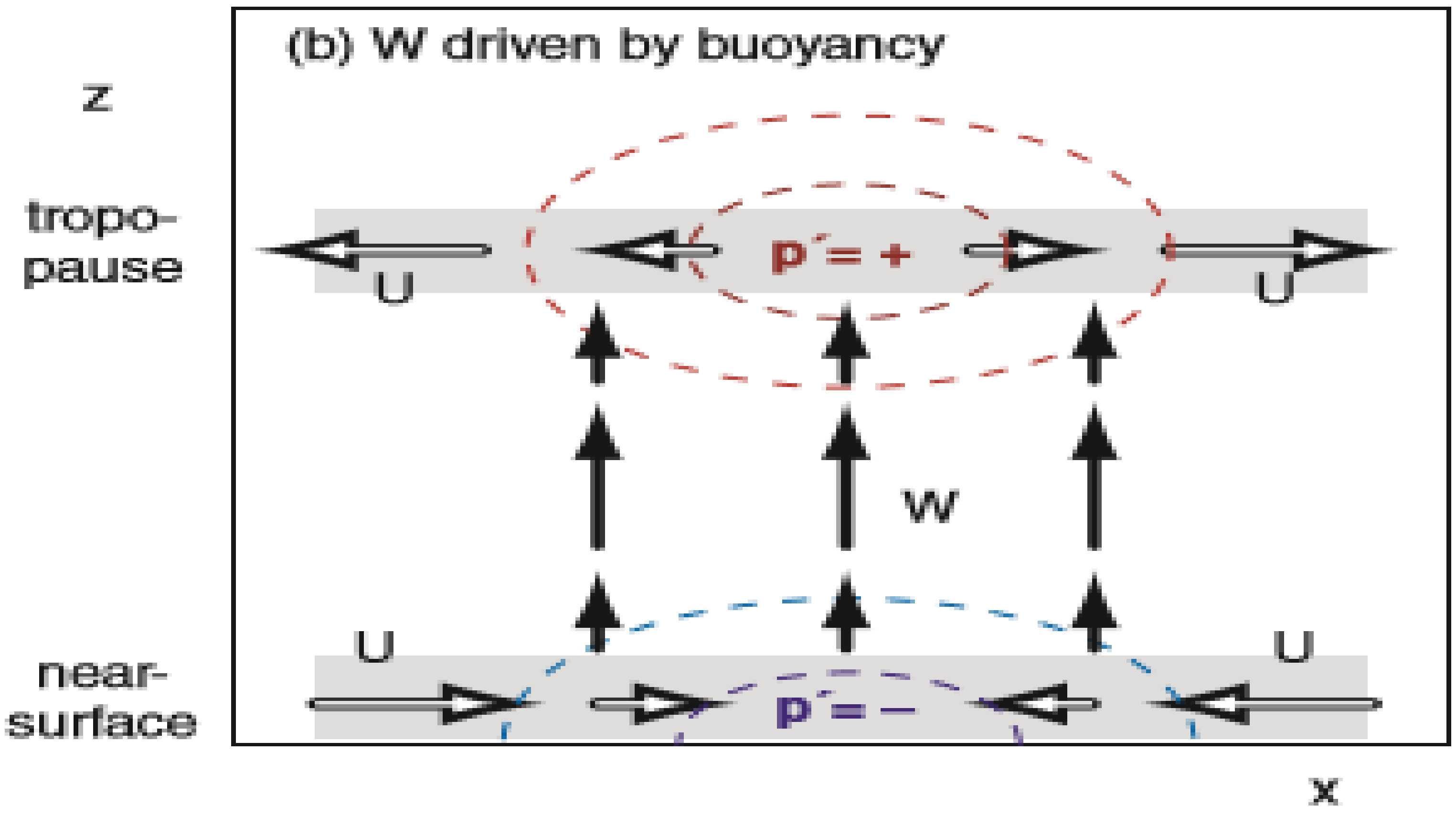
Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

Vertical motions driven by pressure gradients caused by horizontal convergence or divergence.

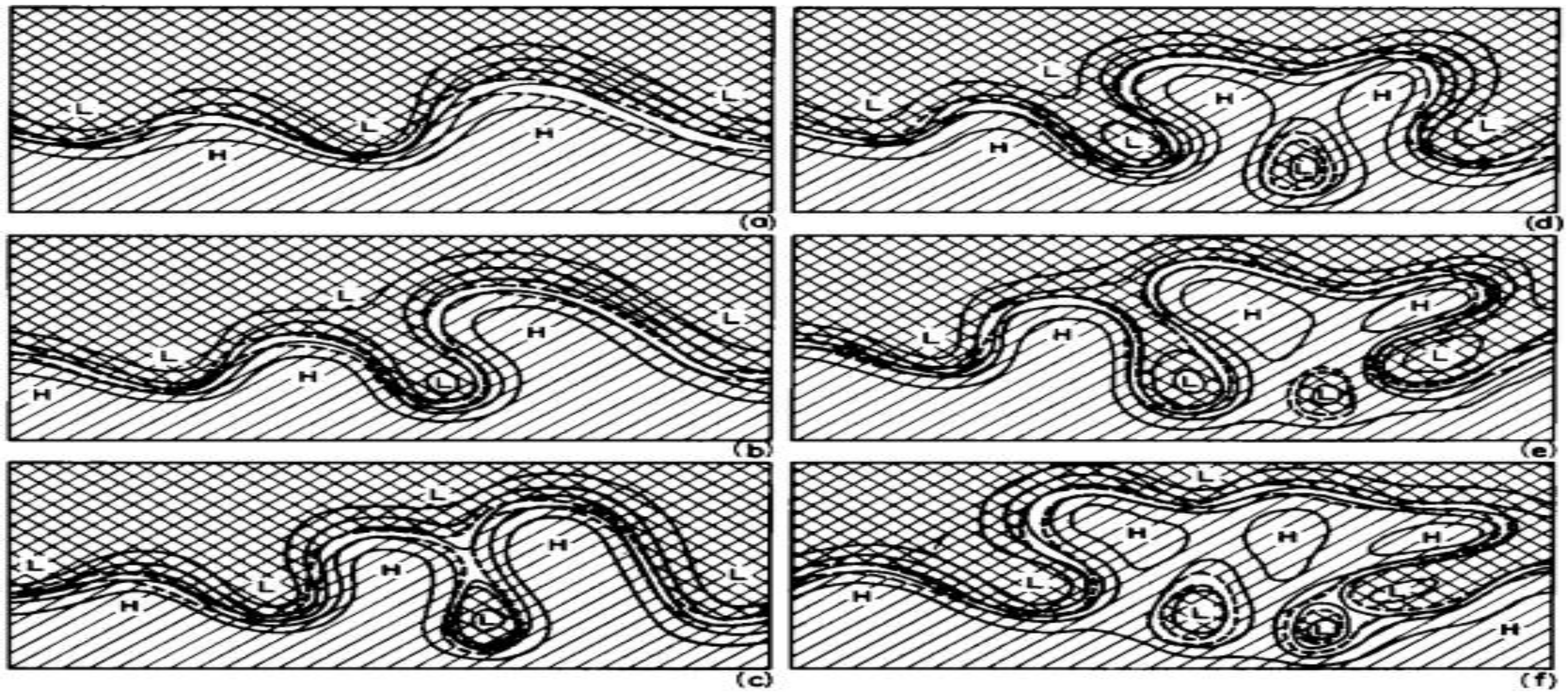


Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

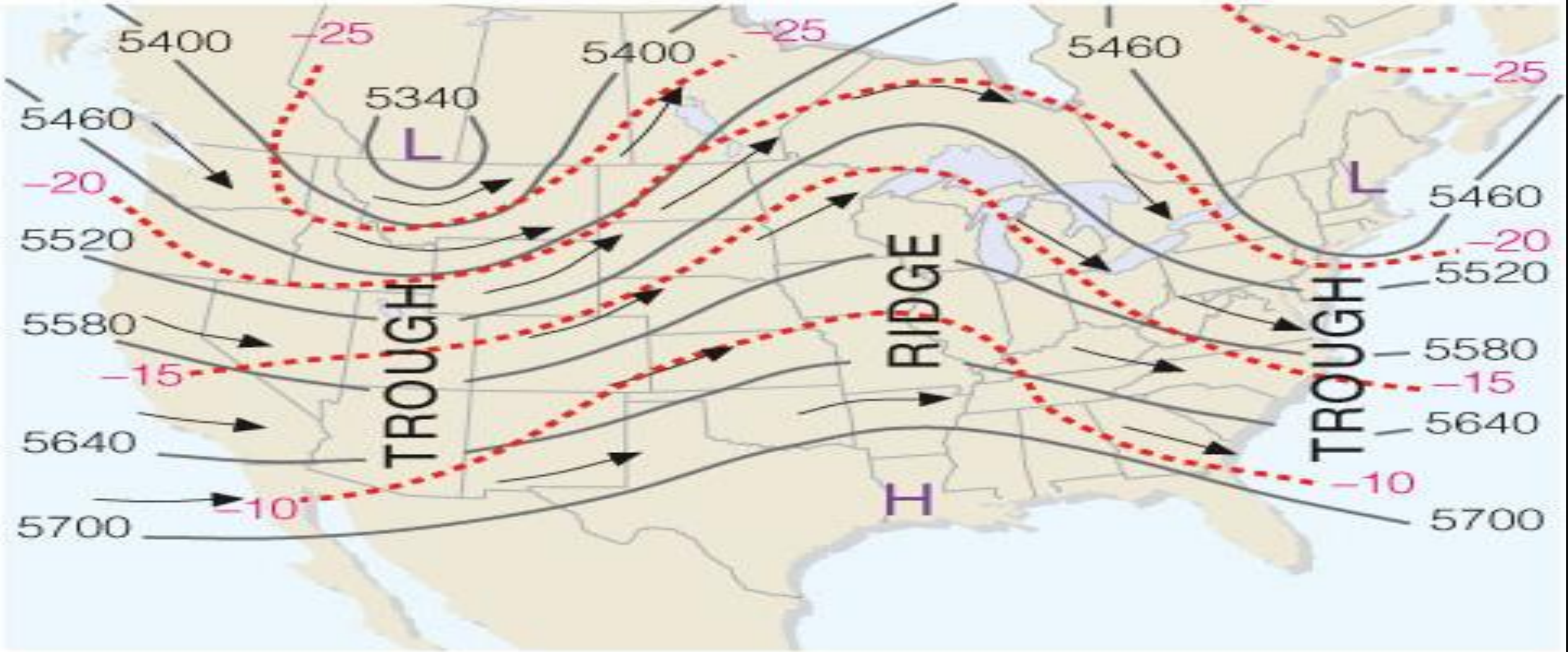
Vertical motions driven by buoyancy.



Chapter 4 General Circulation Upper Atmospheric and Flow Patterns



Chapter 4 General Circulation Upper Atmospheric and Flow Patterns



(b) 500-millibar map

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

The equatorward portions of the wave are known as low-pressure troughs, and poleward portions are known as high-pressure ridges. These ridges and troughs are very transient, and generally shift from west to east relative to the ground.

In regions of straight isobars above the top of the boundary layer and away from the equator, the actual winds are approximately geostrophic.

These winds blow parallel to the isobars or height contours, with low pressure to the left in the Northern Hemisphere

The wind is faster in regions where the isobars are closer together (i.e., where the isobars are tightly packed) and at lower latitudes

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

Balanced Wind Flow

Geostrophic Wind

The geostrophic wind (Ug , Vg) is a theoretical wind that results from a steady-state balance between pressure-gradient force and Coriolis force

$$0 = -\frac{1}{\rho} \cdot \frac{\Delta P}{\Delta x} + f_c \cdot V$$

$$0 = -\frac{1}{\rho} \cdot \frac{\Delta P}{\Delta y} - f_c \cdot U$$

Solving these equations for U and V, and then defining U ≡ Ug and V ≡ Vg , gives:

$$U_g = -\frac{1}{\rho \cdot f_c} \cdot \frac{\Delta P}{\Delta y}$$

$$V_g = +\frac{1}{\rho \cdot f_c} \cdot \frac{\Delta P}{\Delta x}$$

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

The total geostrophic wind speed G is:

$$G = \sqrt{U_g^2 + V_g^2}$$

Gradient Wind

Around a high or low pressure center, the steady state wind follows the curved isobars, with low pressure to the left in the Northern Hemisphere.

Around lows, the wind is slower than geostrophic, called sub-geostrophic, regardless of the hemisphere. Around highs, the steady-state wind is faster than geostrophic, or super-geostrophic.

The curved steady-state wind is called the gradient wind.

The gradient wind occurs because of an imbalance between pressure-gradient (FPG) and Coriolis forces (FCF); namely, the net force (Fnet) is not zero.

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

This net force is called centripetal force, and is what causes the wind to continually change direction as it goes around a circle. By describing this change in direction as causing an apparent force (centrifugal), we can find the steady state gradient wind:

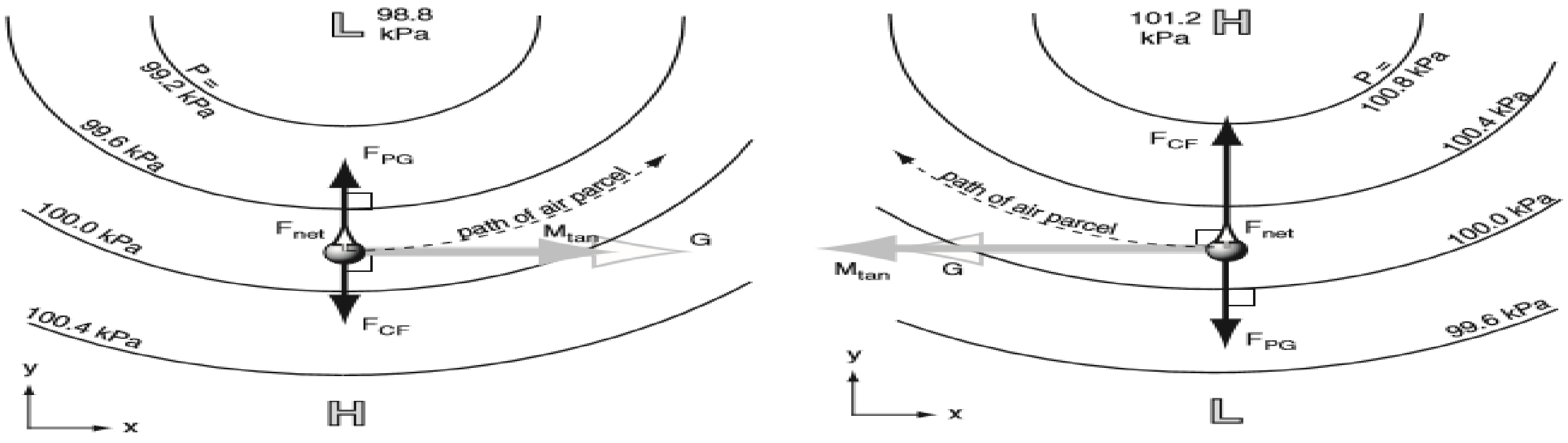
$$0 = -\frac{1}{\rho}\cdot\frac{\Delta P}{\Delta x} \quad + f_c \cdot V \quad + s\cdot\frac{V\cdot M}{R}$$

$$0 = \underbrace{-\frac{1}{\rho}\cdot\frac{\Delta P}{\Delta y}}_{\substack{\text{pressure}\\ \text{gradient}}} \quad \underbrace{- f_c \cdot U}_{\text{Coriolis}} \quad \underbrace{- s\cdot\frac{U\cdot M}{R}}_{\text{centrifugal}}$$

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

Because the gradient wind is for flow around a circle, we can frame the governing equations in radial coordinates: $\frac{1}{\rho} \cdot \frac{\Delta P}{\Delta R} = f_c \cdot M_{tan} + \frac{M_{tan}^2}{R}$

where R is radial distance from the center of the circle, fc is the Coriolis parameter, ρ is air density, ΔP/ΔR is the radial pressure gradient, and M_{tan} is the magnitude of the tangential velocity; namely, the gradient wind.



Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

Cyclostrophic Wind

In intense vortices, strong winds rotate around a very tight circle. Winds in tornadoes are about 100 m/s, and in water spouts are about 50 m/s. As the tornado strengthens and tangential winds increase, centrifugal force increases much more rapidly than Coriolis force. Centrifugal force quickly becomes the dominant force that balances pressure-gradient force

Thus, a steady-state rotating wind is reached at much slower speeds than the gradient wind speed.

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

For steady state winds, the equations of motion reduce to:

$$0 = \underbrace{-\frac{1}{\rho} \cdot \frac{\Delta P}{\Delta x}}_{\text{pressure gradient}} \underbrace{+s \cdot \frac{V \cdot M}{R}}_{\text{centrifugal}}$$

$$0 = -\frac{1}{\rho} \cdot \frac{\Delta P}{\Delta y} - s \cdot \frac{U \cdot M}{R}$$

Cyclostrophic winds never occur around high pressure centers, because the strong pressure gradients needed to drive such winds are not possible.

Around lows, cyclostrophic winds can turn either counterclockwise or clockwise in either hemisphere, because Coriolis force is not a factor.

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

inertial Wind Steady-state inertial motion results from a balance of Coriolis and centrifugal forces:

$$0 = f_c \cdot M_i + \frac{M_i^2}{R}$$

where Mi is inertial wind speed, fc is the Coriolis parameter, and R is the radius of curvature.

Chapter 4 General Circulation Upper Atmospheric and Flow Patterns

Summery

Item	Name of Wind	Forces	Direction	Magnitude	Where Observed
1	**geostrophic**	pressure-gradient, Coriolis	parallel to straight isobars with Low pressure to the wind's left*	faster where isobars are closer together. $G = \left\lvert \frac{1}{\rho \cdot f_c} \cdot \frac{\Delta P}{\Delta d} \right\rvert$	aloft in regions where isobars are nearly straight
2	**gradient**	pressure-gradient, Coriolis, centrifugal	similar to geostrophic wind, but following curved isobars. Clockwise* around Highs, counterclockwise* around Lows.	slower than geostrophic around Lows, faster than geostrophic around Highs	aloft in regions where isobars are curved
3	**boundary layer**	pressure-gradient, Coriolis, drag	similar to geostrophic wind, but crosses isobars at small angle toward Low pressure	slower than geostrophic (i.e., **subgeostrophic**)	near the ground in regions where isobars are nearly straight
4	**boundary-layer gradient**	pressure-gradient, Coriolis, drag, centrifugal	similar to gradient wind, but crosses isobars at small angle toward Low pressure	slower than gradient wind speed	near the ground in regions where isobars are curved
5	**cyclostrophic**	pressure-gradient, centrifugal	either clockwise or counterclockwise around strong vortices of small diameter	stronger for lower pressure in the vortex center	tornadoes, waterspouts (& sometimes in the eye-wall of hurricanes)
6	**inertial**	Coriolis, centrifugal	anticyclonic circular rotation	coasts at constant speed equal to its initial speed	ocean-surface currents

** For Northern Hemisphere. Direction is opposite in Southern Hemisphere. ** Antitriptic winds are unphysical; not listed here.*

End

Next Chaper:5