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The Earth system:
biological and ecological dimensions of global environmental change

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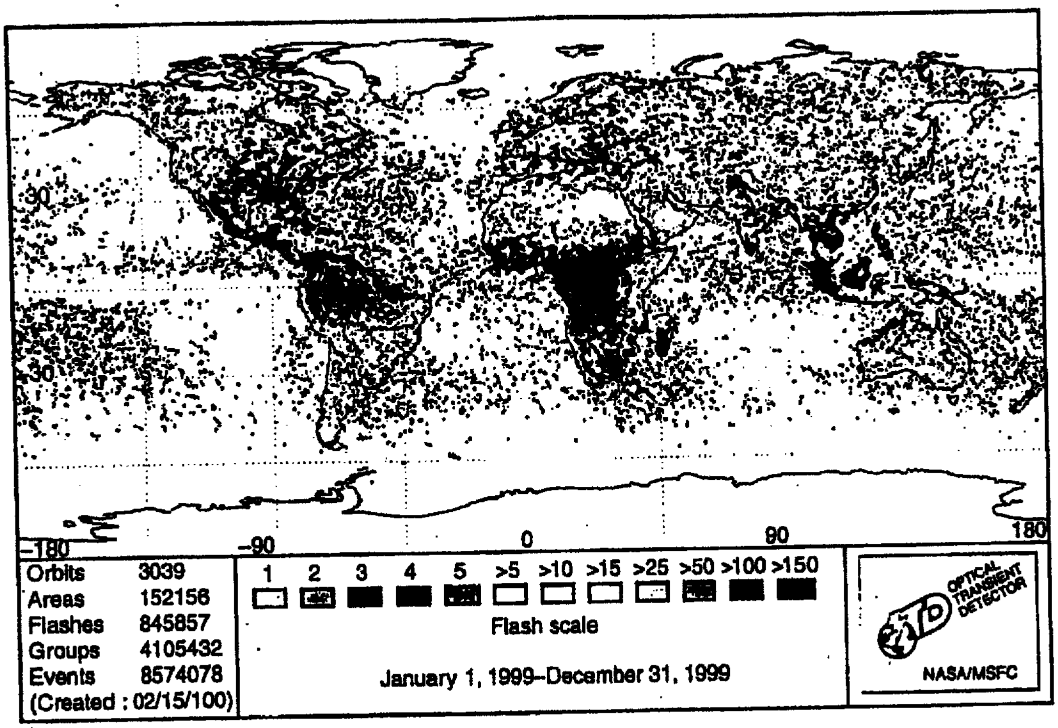

Figure 3 Annual global lightning flashes for 1999, as determined by the NASA satellite-borne optical transient detector. The lightning flash scale indicates the number of flashes per square kilometer per year. Darker regions indicate a greater number of flashes. (Data provided by the Global Hydrology Resource Center at the Global Hydrology and Climate Center, Huntsville, AL (see http://thunder.msfc.nasa.gov/data))

EFFECTS OF GLOBAL CLIMATE CHANGE

The effects of global climate change are uncertain, but one possible implication is a change in the planetary thunderstorm activity. This could lead to a modified lightning distribution, which may have significant feedbacks, because lightning affects the fixation of atmospheric nitrogen, forest fire initiation and ozone concentrations. If climate change does affect lightning, this may indirectly affect the other components of the global atmospheric electrical circuit, although the magnitude and sign of the effect are difficult to determine. For example, the ions comprising the fair weather current flow in the circuit may be relevant in condensation nuclei formation, which may influence cloud formation and properties. The effects of global climate change on such complex and sensitive processes are not yet known.

FURTHER READING

Chalmers, J A (1967) *Atmospheric Electricity*, 2nd edition. Pergamon Press, Oxford.

MacGorman, D R and Rust, W D (1998) *The Electrical Nature of Storms*. Oxford University Press, Oxford.

Rogers, R R and Yau, M K (1989) *A Short Course in Cloud Physics*, 3rd edition. Pergamon, Oxford.

Atmospheric Electricity, Relation to Lightning

see Lightning and Atmospheric Electricity (**Volume 1**)

Atmospheric Model Intercomparison Project (AMIP)

see AMIP (Atmospheric Model Intercomparison Project) (**Volume 1**)

Atmospheric Motions

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With the first pictures looking down on the Earth from a space platform in the 1960s, the beauty and complexity of *the atmospheric motions were immediately evident: the long streaks of jet stream clouds, the large spirals of developing mid-latitude storms, and the strings of pearls of towering thunderstorm clouds in the equatorial regions.*

It was a beautiful view, but it was not entirely unexpected. By the end of the 19th century meteorologists had managed to deduce the essential parts of the global circulation, from soundings by balloons and cloud drift observations at higher levels. Steady improvements of observational technology, radar, remote sensing and, above all satellite observations have provided an ever-increasing amount of data on the state and motion of the atmosphere, not only in the lowest 10–15 km (the troposphere), but also at higher levels (the stratosphere, mesosphere and ionosphere).

There were surprises to come. In the 1890s, when balloon soundings started to reach above 10 km, they recorded that the temperature, from having rapidly decreased, suddenly at some height became more or less constant. This marked the discovery of the tropopause, the cold dividing boundary between the troposphere and the stratosphere. During the First World War, when the thunder on the western front could be heard in London, meteorologists began to speculate that the only explanation must be a warm layer at a height of about 50 km; the existence of the stratopause, the warm boundary between the stratosphere and the mesosphere was later confirmed by more advanced measurements. The third great surprise was of course the jet streams: long, narrow bands of very strong winds of 30 m s^{-1} or more at high altitudes. In the 1960s observations from 30–50 km height in the equatorial region revealed that the winds turned from westerly 20 m s^{-1} to easterly 50 m s^{-1} over a period of, oddly, 13 months. In recent decades the interactions between the oceans and the atmosphere, most notably the El Niño, have added to our understanding of the complexities of the atmosphere.

The main task for meteorology is to explain, or at least describe, the general circulation of the atmosphere and its variations in accordance with the basic principles of fluid dynamics, radiation and thermodynamics. To what extent are the processes that give rise to these complicated circulation systems really understood? The answer depends to a great extent on what is meant by an explanation.

INTRODUCTION

Computer Models

In 1956, Norman A Phillips, an American meteorologist, achieved something remarkable: he fed a computer with the basic facts about the size of the Earth, its gravitational pull, its rotational velocity and the chemical composition of the gases that constitute its atmosphere. Programmed with simplified versions of the mathematical equations that describe the atmosphere's motions and physical processes, the computer was able to generate the observed three-dimensional general circulation of the Earth's atmosphere, including jet streams, travelling cyclones, fronts and high-pressure areas.

Doubling of computer power every one and a half years has since then provided an ever increasing ability to simulate and forecast most atmospheric features. Numerical simulations of the general circulation using supercomputers serve as an atmospheric laboratory, in which we can perform experiments, observe the effects and hopefully predict its future state. However, many meteorologists are not satisfied and think that it is not enough that the "computers seem to understand." We want to know, for example, why the westerly winds so dominate the mid-latitudes, why the Sun's radiation reaches the Earth's surface without being hindered by clouds in the subtropical areas, why northwestern Europe's winters are so mild compared to those at the same latitude in northeastern North America and why the storms in the tropics are so different from the storms that rage over the Atlantic or the Pacific.

Watching Atmospheric Motion

Our understanding of the atmosphere is like that of the spectator at a football match. Without any insight into the rules of the game, it will only appear to be a lot of people randomly running around chasing a ball. Those who know the rules and the logic of the game might not be able to make predictions of what will happen in the next minute, but can make sense of what is unfolding in front of their eyes.

Our knowledge of the atmosphere is often summarized in a statistical way. But using statistics brings advantages and disadvantages. We gain in simplicity, but also lose because we have thrown something away. Whenever there are averages, there are fluctuations or deviations from these averages. In addition, a mean picture does not necessarily represent a typical picture. In environmental studies, we are for example interested to know the evolution and trajectory of individual air parcels: Where does the moist air go? From where comes the polluted water? What makes the air lose its ozone? One good illustration of how knowledge can enhance understanding is the movement of a water droplet in a simple water wave.

Two Ways to Look at Motion

Consider a train of waves on a water surface moving from left to right, as depicted in Figure 1(a). The fluid is rising as the ridge of the wave approaches and falling after it passes. The mean motion, as inferred by averaging

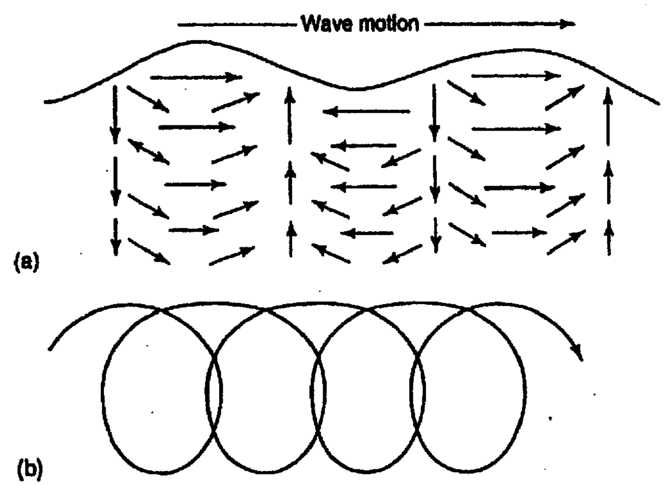


Figure 1 (a) A train of waves moving from left to right. For further discussion, see text. (b) The individual particles in the water are subjected to a net displacement to the right

the motions at individual points in space over a large number of waves, will everywhere be zero because all motions, vertical and horizontal in the long run cancel out (sometimes called *Eulerian averaging*). However, if we concentrate on individual parcels, we can see that they move in clockwise circuits. If we look carefully, we see that the extent of the motion decreases with depth, so we may realize that parcels move a little farther towards the right at the top of their path, than they move toward the left farther down. Hence, there is a net rectifying displacement toward the right, as pictured in Figure 1(b). To understand the motions of the air we must therefore also make use of averages that are geared to the movement of individual air parcels (sometimes called *Lagrangian averaging*).

Motion without Fricton

The motions of the atmosphere show patterns, regularities and mutual relationships, which make sense if certain basic mechanisms are understood. Many of these patterns are easily accommodated with everyday experience: the bubbles in a pan of boiling water resemble the production of convective clouds; the downhill flow of water resembles the katabatic winds that flow down mountain slopes; etc. What is difficult is the seemingly simple. Many processes, particularly in the free atmosphere, occur under almost frictionless conditions. Pure inertia is actually a very difficult process to comprehend with our senses because our everyday life very much involves friction. The fact that the motion takes place on a rotating planet also has some counterintuitive consequences.

These energy conversions occur in the interplay between gravity, the effect of the Earth's rotation and friction.

THE DRIVING AND RETARDING FORCES

The Sun provides the energy that drives the atmospheric circulation, but this energy is not evenly distributed over the globe. While gravity acts to equalize these differences by setting the air in motion, friction seeks to bring all motion ultimately to rest.

The Incoming Radiation from the Sun

Understanding the atmosphere involves the unraveling of a lot of paradoxes. It is, for example, not primarily the Sun that warms the atmosphere, but the Earth. Or rather the Sun warms the air indirectly by first warming the surface of the Earth, which then warms the air. The radiation from the Sun is short wave (visible) and the cloud-free atmosphere allows much of it to pass through without much absorption. Of all the solar energy that initially enters the Earth's atmosphere. about half reaches the surface to warm the land masses and oceans. About 30% of incoming solar radiation is reflected back to space, mostly by clouds, snow and ice; and about 20% is absorbed in the atmosphere (*see* **Energy Balance and Climate**, Volume 1).

When the land and oceans heat up, they emit long-wave (infrared) radiation, to which the atmosphere is not very transparent. The atmospheric absorption of long-wave radiation is due to the presence of radiatively active gases, the most important of which is water vapor. Together with carbon dioxide, ozone and the other trace gases (methane, nitrous oxides. chlorofluorocarbons, etc.), these gases account for the normal greenhouse effect, without which most parts of the Earth would be uninhabitably cold because much more radiation would leave the atmosphere and much less downward long-wave radiation would warm the surface (*see* **Infrared Radiation**, Volume 1). Although there is a debate about the magnitude of the *enhanced* greenhouse effect caused by human emissions, there is no such debate about the greenhouse effect as such (*see* **Greenhouse Effect**, Volume 1). What further complicates this already complex balance of in- and out-going radiation is that it depends on latitude, the seasons and the distribution of land and sea.

The Uneven Heating and Cooling

The farther away from the tropics, the lower the Sun rises over the horizon. This means that the Earth's surface is not as efficiently heated in the higher latitude regions as in lower latitudes. The incoming solar radiation depends also on the orientation of the hemispheres relative to the Sun; thus it changes with the seasons. The summer hemisphere receives more insolation than the winter hemisphere; during the summer the polar regions receive as much solar radiation per day as the tropics!

The outgoing radiation is, however, less dependent on latitude. Because land and sea have different physical properties, they react differently to the influx of solar radiation. Land has a comparatively small heat capacity and its immobility means that heat can only spread through conduction and diffusion. Land surfaces can therefore go through temperature changes on the order of several degrees within hours down to depths of several centimeters. Water bodies have a much greater heat capacity and a lower albedo (reflectivity) than land; water's mobility also enables an efficient transfer of heat vertically as well as horizontally. For these reasons, oceans absorb more solar radiation and to greater depth than land. The vertical mixing in the ocean spreads the heat deeper, so that the sea surface temperature changes little even on seasonal time-scales (*see* **Sea Surface Temperature**, Volume 1). In the latitude band up to 40° north and south the continents are on average warmer than the oceans; poleward of 40° the oceans are warmer. These differences in heating and cooling create differences in the density and the mass distribution of air. It is these contrasts that gravity tries to equalize.

Gravity – The Great Equalizer

If the ultimate source of energy is the Sun, the ultimate mover is the Earth's gravitational pull. Gravitation is always trying to turn everything into perfect spheres, including the Earth and its atmosphere. Due to the centrifugal effect of the Earth's rotation, its shape is a sort of oblate ellipsoid, bulging outwards at the equator (this shape is referred to as the *geoid*). Any material surface that does not follow such a geoid, but stands out against it is subject to a gravitational pull.

Pressure falls off more slowly with altitude in warm air than it does in cold air. Therefore, the same atmospheric pressure is found at higher elevations in the tropics than in the polar region. If the pressure at a certain level is not horizontally uniform, gravity accelerates the air from high to low pressure to equalize the pressure distribution. We talk about this acceleration as due to the *pressure gradient force* (PGF), which is not any new kind of force, just the name we have given to the effects of gravity in this context.

Gravity also generates vertical motions by making colder and heavier air sink and warmer and lighter air rise. Rising air experiences decreasing pressure with height, and so expands and cools; sinking air encounters increasing pressure, and so is compressed and warms. Rising air therefore tends to have a cooling effect on the environment, sinking air to have a warming effect. One important exception is, as we will see, rising moist air where the water vapor condenses.

The Three Phases of Water

Heat melts ice into water, and vaporizes water into a gas. The energy used for these processes is stored in liquid water and water vapor, respectively. When water vapor condenses into liquid water, or water freezes into ice, this *latent heat* is released again. A rainfall of 8 mm, which is the same as 8 liters per square meter, releases the same amount of energy as the same area would have received from the Sun during one day. The same happens during snowfall when liquid water freezes to ice, although the amount of energy released is less. A large proportion of the energy received from the Sun is stored in this way and this plays an important role in the dynamics of the Earth's atmosphere because energy can be transported long distances before it is released as latent heat.

Ice, water and water vapor are important also for their reflective characteristics. The proportion of solar radiation that is reflected back to space (albedo) varies significantly with location because it depends on the proportion of ice-covered land and sea, and on the ice crystals in the upper parts of the clouds. Water vapor also absorbs and emits radiation differently from liquid water. The water drops and ice crystals in the clouds reflect the Sun's short wave radiation, which prevents it from reaching the Earth's surface. The expected cooling of the air beneath the cloud is to some extent compensated by warming from long-wave radiation that is trapped by the same cloud.

Finally, water has a high specific heat capacity; that is, water takes a lot of energy to warm, and it cools slowly. Energy stored in the oceans at tropical latitudes does not dissipate rapidly when transported to higher latitudes, serving as a regulator and conveyor of energy. Without these ocean currents, the temperature difference between poles and equator would be 90–100 K instead of the 40–45 K that is observed. The energy conversions related to water in all three forms, and their different radiative properties, contribute to making predictions of any change in the climate very difficult.

Friction

Friction plays a double role: on the one hand it is slowing down all motion; on the other hand, it is through friction (and gravity) that the atmosphere feels that it is on a rotating Earth. Large-scale friction like lee-effects of the large mountain ranges affects the atmosphere's tendency to form large-scale vortices.

THE RELATIONSHIP BETWEEN WIND AND PRESSURE

When the PGF starts to equalize by accelerating the air from high to low pressures, the rotation of the Earth tries

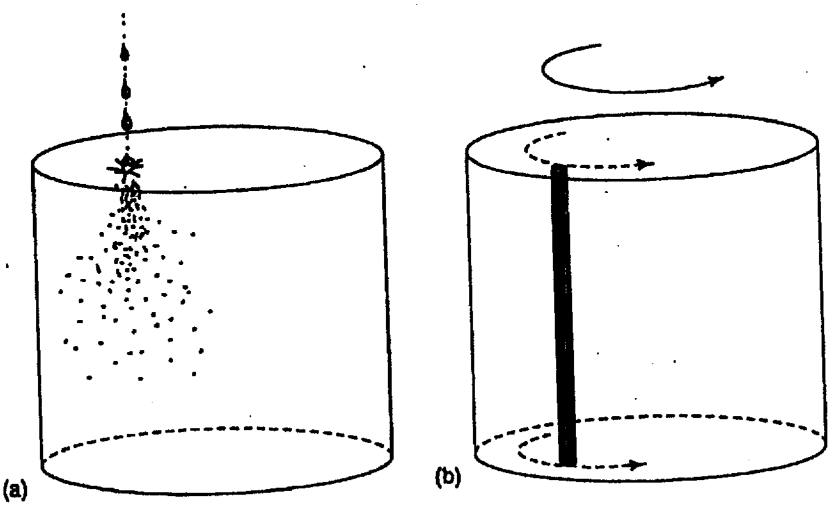


Figure 2 (a) Ink inserted into a non-rotating water tank would disperse immediately. (b) The same ink inserted into a rotating tank would, due to the Coriolis effect, form vertical columns. Such vertical columns are called "Taylor columns" after the British scientist who first conducted this experiment

to return the air from where it came. The struggle between these two contesting forces shapes the pressure and wind patterns.

The Effect of the Rotation of the Earth on an Air Parcel

Due to the Earth's rotation, wind and ocean currents are deflected to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. This is due to the so-called *Coriolis force* (fV, where f is the Coriolis parameter and V the wind speed) (*see* **Coriolis Effect**, Volume 1); it has the effect of opposing any displacement by trying to restore the air to its initial position.

A striking demonstration of this can be made using a rotating tank of water. When a drop of ink is inserted in the water, it falls slowly downward. But instead of dispersing and coloring the water, it remains in a vertical column moving around with the tank as a rigid body. What happens is that when the ink particles start to spread out horizontally, they are immediately affected by the Coriolis force (fV), which forces them into circular motions. If the tank rotates with one revolution in two seconds ($\omega = 2\pi/2 = 3.14\,\mathrm{rad\,s^{-2}}$), and the ink spreads out with a velocity of $1\,\mathrm{cm\,s^{-1}}$, the fV would yield circles with radius of less than 2 mm (Figure 2).

An airflow in the atmosphere of $10\,\mathrm{m\,s^{-1}}$, only affected by the fV, would be confined in a circular trajectory with a radius of 100 km, about the size of London or Los Angeles. In the equatorial regions, where the horizontal component of the fV is relatively weak, velocities of the order of $30\,\mathrm{m\,s^{-1}}$ would be needed to move an air parcel from the equator to poleward of 20°latitude. Were it not for the existence of strong PGFs, it would be very difficult for air parcels in the atmosphere to move any considerable distance.

The Relationship Between the Wind and Pressure

With two effects, the PGF and the fV, pulling in opposite directions, it is no surprise that they quite often balance each other,

$$f \cdot V = \mathrm{PGF}$$

If this is the case, the air moves by itself with no acceleration (Figure 3). Winds blowing under these conditions are called *geostrophic* (from Greek *geos* meaning Earth and *strophein* meaning turning). For such conditions,

$$V_{\mathrm{geostrophic}} = \frac{\mathrm{PGF}}{f}$$

The farther away from the equator, the stronger the PGF has to be to balance the fV. That is why pressure maps over the tropics and subtropics contain relatively

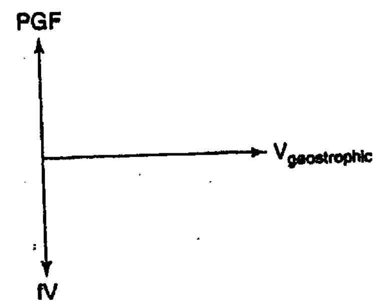


Figure 3 The wind is geostrophic if the PGF and the fV, balance each other

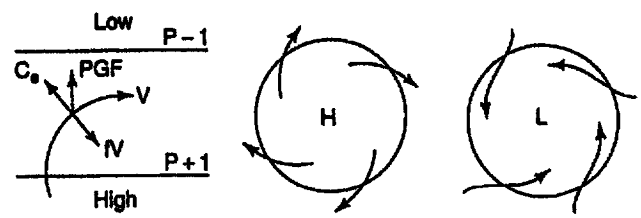


Figure 4 An air parcel accelerating into a pressure field will describe a curved motion and be affected also by a centrifugal force (C_e). A wind accelerated out from a high pressure area by a PGF will be deflected to the right by the fV and, after adjustment describe a clockwise circulation; a wind accelerated into a low pressure area by PGF will be deflected to the right by the fV and, after adjustment describe an anti-clockwise circulation

fewer isobars (lines of equal pressure) than at higher latitudes.

The Effect of Curved Flow

When the pressure field drives the air into a curved motion, it is affected also by a *centrifugal force* (C_e) that disrupts the bilateral balance between the fV and the PGF (Figure 4). Centrifugal forces are always directed outward from the center of rotation and, depending on how the trajectory is curved, the C_e will support either the fV or the PGF (Figure 5).

In a stationary high-pressure system (where the air moves in clockwise trajectories), the C_e supports the PGF; in a stationary low-pressure system (where the air follows anticlockwise trajectories), the C_e supports the fV. This is why high pressure systems, where the C_e supports the PGF, normally have weaker pressure gradients (fewer isobars); and low pressure systems, where the PGF must balance both the fV and the C_e, generally have stronger pressure gradients (more isobars).

Airflow in which the PGF, the fV and the C_e balance each other is called *gradient flow*. In small, intense vortices, with high velocities and small radii, such as in tropical cyclones, the C_e may dominate over the fV. These are called *cyclostrophic flows* (Figure 6).

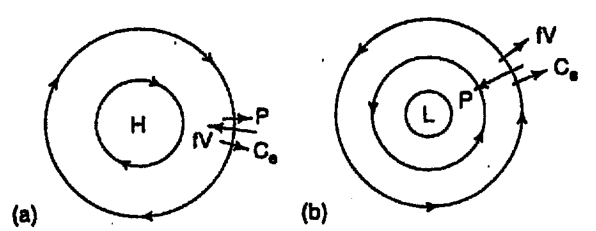


Figure 5 (a) The gradient wind approximation for cyclonic and anticyclonic flows. For curved air trajectories around a high pressure area (H), the C_e will support the PGF P; (b) for curved trajectories around a low pressure area (L), the C_e will support the fV

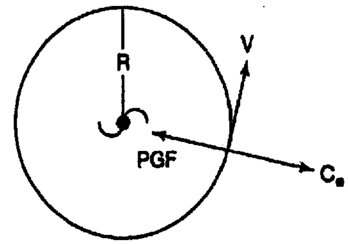


Figure 6 For very intense vorticies like mature tropical cyclones, the C_e dominates over the fV, which can be neglected, although it may have played a decisive role in the initial stages of the formation

The Effect of a Moving Flow Pattern

If the circulating system is not stationary, but moving, the relation between the C_e, the PGF and the fV is more complicated. The trajectories will alternately be more or less curved, compared to those of a stationary vortex. The reason why the winds on the southern side of a cyclone moving eastward close to geostrophic is because the air trajectories are rather straight (a large radius of curvature). On the other hand, on the northern side, the winds are much weaker, although the pressure gradients might be the same, because the air trajectories are more curved (Figure 7).

The Dynamics of Pressure Changes

Had the wind in the free atmosphere been in complete geostrophic, gradient or cyclostrophic wind balance, it would be frozen in a permanent state. What instead keeps the atmospheric circulation changing is that these balances are never fulfilled. It is the *ageostrophic* wind component, the part of the wind that is not in geostrophic balance, that drives the atmospheric patterns. If the wind at some instant happens to be in geostrophic balance, it sooner or later moves into a region where it is no longer in geostrophic balance (Figure 8). The PGF and the fV then provide a mechanism for restoring the balance by moving mass horizontally between lower and higher pressure.

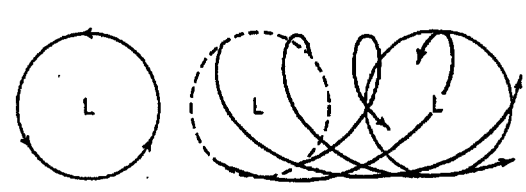


Figure 7 For an eastward moving circular vortex the trajectory of an air parcel on the southern side (where the air is moving in the same direction as the vortex) will have less curved trajectory than on the opposite side, where the air is moving in the opposite direction the trajectory will be more curved

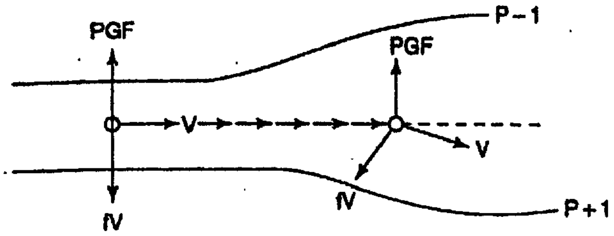


Figure 8 The wind is only temporarily in geostrophic balance; the motion of the air soon brings it into areas where this balance is no longer valid. In the schematic picture the wind is in geostrophic balance upstream where the pressure gradient is stronger than further downstream. When the wind enters this downstream region, it becomes supergeostrophic, and is deflected to the right by fV. The flow will thereby transport air towards higher pressure, which will strengthen the pressure gradient at the same time as the wind weakens it, until a new balance is reached

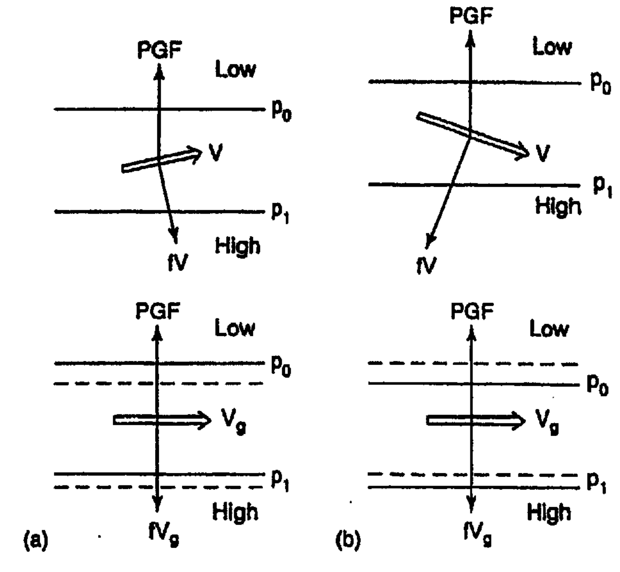


Figure 9 (a) When the PGF is stronger than the fV the wind *accelerates* towards lower pressure. At the same time the PGF weakens when mass is transported from high to low pressure. The wind moves parallel to the lines of equal pressure (isobars) when the two forces fV and PGF have reached a balance. (b) When the PGF is weaker than the fV the wind *decelerates* towards higher pressure. At the same time the PGF strengthens when mass is transported from low to high pressure. The wind moves parallel to the isobars when the two forces fV and PGF have reached a balance

If the wind is sub-geostrophic, i.e., weaker than the geostrophic balance requires, the PGF will accelerate the air towards lower pressure. Mass will be transported from higher to lower pressure and thereby weaken the PGF. At the same time the wind will increase, and so will the fV, until a new temporary balance with the PGF is established (Figure 9a).

If the wind is supergeostrophic, i.e., stronger than the geostrophic balance requires, the fV will drive air towards higher pressure. Doing so, an air mass will be transported from low to high pressure, thereby increasing the PGF. At the same time, the wind will weakens and so will the fV, until a new temporary balance with the PGF is established (Figure 9b).

The Ice Skater Effect

An important mechanical effect in the atmosphere is the *ice skater* effect. By contracting or widening their arms, ice skaters increase or decrease their rate of rotation. Something similar happens for a body of gas (or fluid) which contracts or expands. But there are also vertical motions in the gas. Rising motion is associated with an inflow of air at lower layers and an outflow in upper layers. Sinking motion is associated with an inflow of air at upper layers and an outflow in lower layers. Due to the fV, the air flowing inwards forms an anticlockwise spiral; air flowing outwards forms a clockwise spiral. The pressure then adjusts to create a low pressure distribution for the inward spiral or a high pressure distribution for the outward spiral (Figure 10).

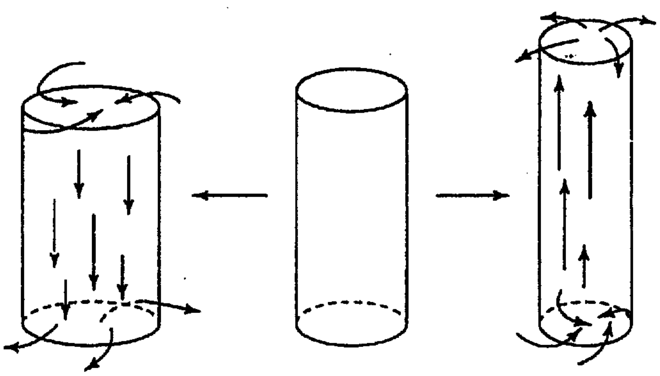


Figure 10 Schematic illustration of the ice skater effect. In the Northern Hemisphere, upward motion is associated with air spiraling anticlockwise inward at lower levels, clockwise outward at higher. For downward motion the opposite is the case

THE ATMOSPHERIC ENGINE

On average the equatorial region receives more heat from the Sun than is sent back to space, while the opposite is true at high latitudes. The accumulated result of the heating is an expansion of the air at low latitudes, setting the atmospheric engine in motion.

Hadley Circulation

Because pressure falls off more slowly with height in warm air than in cold air, the same pressure in the atmosphere is

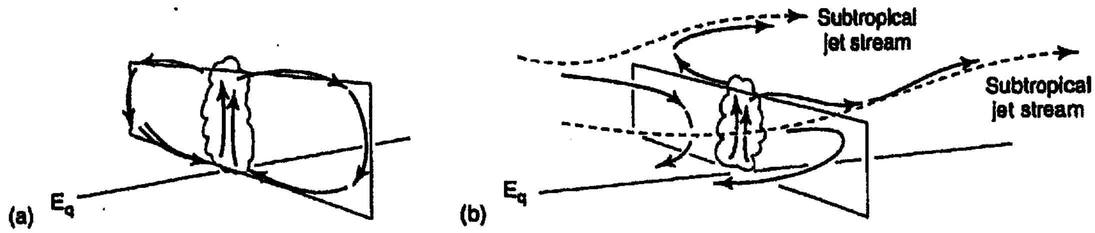


Figure 11 The Hadley cell is characterized by rising motion in the equatorial region, poleward flow at upper levels, sinking at higher latitudes and a return flow equatorward at low levels. (a) The traditional Eulerian view, which depicts the average north–south motion and vertical motion; (b) an attempt at a Lagrangian view, which aims at depicting the average motion of individual air parcels which necessarily involves a three dimensional image including and east–west motion

found at higher elevations in the tropics than elsewhere. From these upper levels, air is accelerated towards the poles. This is the driving mechanism behind the Hadley circulation, named after the 18th century British scientist who first suggested the Earth's rotation and differential heating were major components in the atmosphere's general circulation (Figure 11). The removal of air from the upper levels is felt farther down as a decrease in pressure and leads to a groove of low pressure at low levels along the Equator. From the surrounding areas of relatively higher surface pressure, air is accelerated into this groove, creating the *Trade Winds*. Due to the fV, the Trade Winds arrive at the equator from a northeasterly direction in the Northern Hemisphere and from a southeasterly direction in the Southern Hemisphere (*see* **Hadley Circulation**, Volume 1).

Intertropical Convergence Zone.

A large part of the heating in the equatorial region does not originate directly from the Sun, but through a delayed effect. Through evaporation from rivers, seas and the vegetation the Sun's energy is stored in the atmosphere as water vapor. It is then transported by the Trade Winds and converges in a band along the equatorial zone called the Intertropical Convergence Zone (ITCZ). The ITCZ is a surprisingly narrow belt of vigorous convective activity and heavy precipitation, usually made up of a large number of distinct cloud clusters a few hundred kilometers wide separated by clear skies. Of particular importance are the so-called *hot towers*, large cumulonimbus clouds that concentrate the main heating.

Over the oceans the ITCZ is not located exactly at the equator. Sometimes it is split up into two parallel bands on either side of the equator. This is because of the relatively low sea surface temperature often present at the equator. How this colder water has been brought there is worth a short explanation because it involves the action of the fV. The steady Trade Winds from the east drive the surface water towards the west. Although the fV is weak close to the equator, it has an accumulated deflective effect on the water due to the steadiness of the driving wind, and so fV deflects the surface water away from both sides of the equator. This leads to an outflow of surface water, which causes an upwelling of deeper and colder water from deeper layers. This mechanism is particularly prominent over the eastern Pacific Ocean where an equatorial pool of cold water generally prevails. This is known as La Niña (Figure 12) (*see* **Intertropical Convergence Zone (ITCZ)**, Volume 1).

Monsoons

The fact that land has a small heat capacity compared to ocean is profoundly significant for many atmospheric features. The absorption of solar radiation raises the surface temperature over the land much more rapidly than over the oceans and causes and creates local wind systems. In the smallest scale we find the sea breeze, in the largest scale the monsoon, a gigantic land and sea-breeze circulation caused by the seasonal warming and cooling.

The largest monsoon circulation is created by the vast Asian landmasses and the Himalayas. During the cold season, the cooling of the Asian continent north of the Himalayas causes cold air to sink. Sinking motion is linked with inflow at upper levels and outflow at lower levels, the ice skater effect. The fV drives these flows into outward spiraling anticyclonic and inward spiraling cyclonic patterns, respectively. Due to the pressure adjustment, a vigorous surface high-pressure area is created over Siberia, with a tendency to lower pressure aloft. The cold air from Siberia moves southward over China and Southeast Asia. This is the winter monsoon.

During the warm season the heating creates a vast upper air high-pressure system that is centered over the Indian peninsula and the Himalayan mountain plateau. This leads to an upper outflow; a surface low-pressure area is formed and moist air is drawn in from the surrounding areas. The weakness of the fV over the Indian Ocean allows air, even from south of the equator, to be drawn into the low-pressure region over the Indian Peninsula. The influx of moisture gives rise to the formation of strong

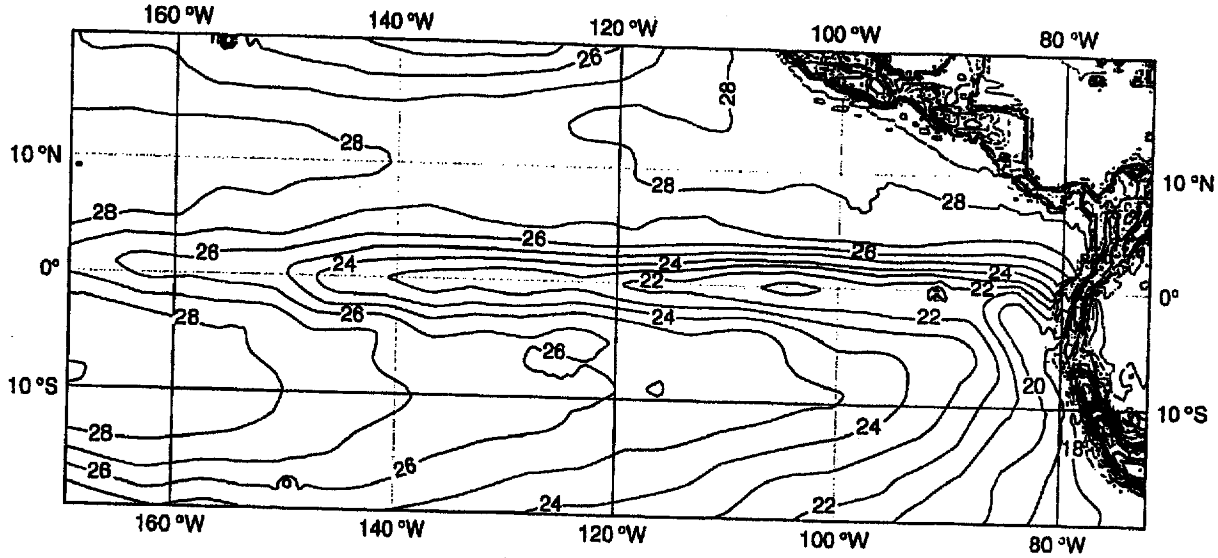


Figure 12 Mean sea surface temperature of the eastern part of the Pacific Ocean for the period 5 Sep to 5 Oct, 1998. The relatively colder water along the equator has been brought up from deeper layers through the action of the fV. Although the fV is weak close to the equator, the steady trade winds from the east account for a small accumulated effect, so that water on both sides of the equator is deflected away from it. This divergence causes an upwelling of deeper and colder water

cumulonimbus convection that leads to further heating of the air due to the release of latent heat.

The monsoon circulation, in particular the summer monsoon, also has an impact on the neighboring continents. The dry regions over northeast Africa are assumed to have their climate determined by the intensity of the Indian monsoon. One contributing factor to this westward directed influence might be the formation of a strong easterly jet stream formed over the northern part of the Indian Ocean on the southern flank of the upper air high pressure system over the Indian peninsula mentioned above (*see* **Monsoons**, Volume 1).

Easterly Waves and Tropical Cyclones

The Trade Winds blow steadily, interrupted only by minor westward moving easterly waves. When they interact with disturbances that have come down from higher latitudes, they develop into subtropical storms, which can become very intense due to the release of latent heat. These disturbances can sometimes move into higher latitudes and affect developments there.

During the warm season, when the subtropical high-pressure belts are displaced poleward, the easterly waves appear in regions with slightly stronger fV. This allows individual easterly waves to develop into small, intense tropical cyclones, hurricanes (or typhoons as they are called over the Pacific, Willy–willys in Australia) driven by the release of latent heat (*see* **Hurricanes, Typhoons and other Tropical Storms – Descriptive Overview**, Volume 1; **Hurricanes, Typhoons and other Tropical Storms – Dynamics and Intensity**, Volume 1). The fV, which has been instrumental during the formation of the initial vortex, soon becomes overpowered by the C_e, resulting in a cyclostrophic flow. The outward pushing C_e causes a sucking effect that draws down dry air from high levels down into the center of the storm. This is the cloudless eye of the storm.

Because tropical cyclones depend crucially on latent heat release from moist air, their development occurs almost exclusively during the peak of the warm season. The air and water temperature (which peak at different times) should not be below +28 °C. Most tropical cyclones die out over land due to friction and lack of supply of sufficiently warm and moist air from below. Because of the cold ocean water from the Antarctic, there are never any hurricanes in the South Atlantic and Southeast Pacific.

Walker Circulation

A third major tropical circulation system driven by heating from below is the so-called *Walker circulation*. Its rising branch is over the Indonesian archipelago and the western part of the equatorial Pacific Ocean, a region of very warm sea surface temperatures and high mountains with intense rainfall, causing extensive release of latent heat. The descending branch is over the central parts of the Pacific where the sea surface is relatively cooler (Figure 13).

Unlike the Hadley circulation and the Monsoon, the Walker circulation is hardly affected by the fV. The air that has been brought to high altitudes over Indonesia is blown straight eastward toward the eastern Pacific. The return flow over the equatorial Pacific is associated with a

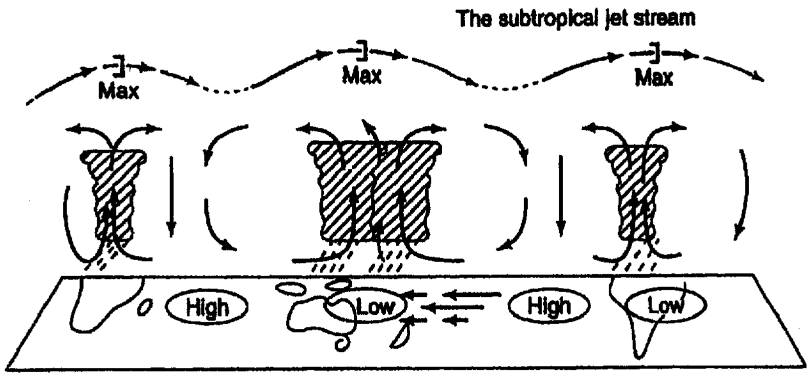


Figure 13 The Walker circulation has rising motion due to convection over Africa, the Indonesian archipelago and South America. The surface flow over the Pacific is easterly, which creates an upslope in the water north of Australia. Shown schematically above these cells are the approximate locations and wind velocity tendencies of the subtropical jet streams that surrounds the Walker circulation at about 30°N and 30°S. With the long distance between the center of the subtropical jet over Indonesia and the center over South America, the Coriolis fV might manage to deflect the jet stream from both hemispheres to cross the equator

weak pressure gradient, with high pressure over the Pacific and low pressure over Indonesia. For not quite understood reasons, this gradient reverses its direction with a period of 2–5 years. These semi-regular variations are known as the Southern Oscillation (*see* **Southern Oscillation**, Volume 1), which is associated with one of the most important semi-regular variations in the ocean, the El Niño (*see* **Walker Circulation**, Volume 1).

El Niño and La Niña

The steady easterly Trade Winds coming from the Pacific into the Indonesian region cause a pile-up of warm surface water, creating an upslope hill. From time to time the Trade Winds weaken, allowing the water to spread back towards the central Pacific. This is the El Niño, which is Spanish for "the Christ child." This term was originally applied to a warming of the coastal waters of Peru and Ecuador that occurs around Christmas time. As the warmest water moves eastward, the tropical convective clouds and rain and heating maximum follows suit, moving towards the central Pacific. In this way, the whole Walker circulation is displaced eastward, which leads to profound consequences throughout the tropics, with changed patterns of rainfall leading to floods in some regions and drought in others (Figure 14).

This shift extends its influence over North and South America, and probably also to the Atlantic where the frequency of tropical storms (hurricanes) is affected. These changes constitute a striking example of how feedbacks between the atmospheric and ocean circulations act together to drive irregular or semi-regular variations, influencing climate and climate change (*see* **El Niño and La Niña: Causes and Global Consequences**, Volume 1).

Because the El Niño (EN) and Southern Oscillations (SO) are highly interrelated phenomena, they are often treated together under the name ENSO. It is not quite known what triggers the change. It could be a change in either the atmospheric or the ocean component of the system. Some observations point to an influence of the decaying monsoon over the Indian Ocean. Still, the ENSO events are becoming predictable, at least to some extent, by computer models. Also, to some extent, their anomalous weather consequences for other parts of the tropics are also predictable, including some aspects affecting North and South America and the Atlantic. An important question is whether the El Niño would have the same character in a warmer atmosphere resulting from enhanced greenhouse warming.

The Formation of the Subtropical Jet Stream

Why does the Hadley circulation extend only to 30° and not to the poles? Part of the answer is contained in an elementary calculation (often found in old meteorological textbooks). The student is invited to calculate the poleward acceleration of air due to the meridional pressure difference at upper levels. A thermal contrast of 40 K is assumed between pole and equator. It turns out that the corresponding acceleration is only $0.7\,\mathrm{mm\,s^{-2}}$. This does not sound like much, but in 24 hours this tiny acceleration would have increased wind speeds to velocities of $60\,\mathrm{m\,s^{-1}}$ and carried the air a distance of more than 2500 km, which is more than 20 degrees latitude.

The accelerated air is increasingly affected by the fV, which tries to make it return to the equator. A balance is eventually established when the air on the one hand is prevented from moving farther poleward by the fV, and on the other is prevented from returning equatorward by a strengthened PGF. With the typical heating and rotation of our particular planet, this balance between wind and

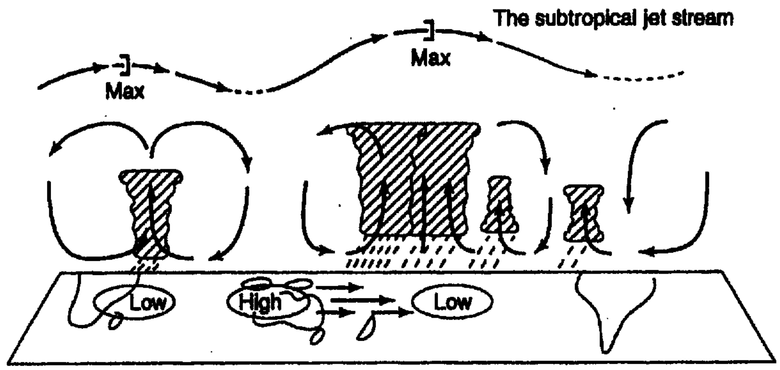


Figure 14 The El Niño occurs when the warm ocean water in the western Pacific spreads back towards the east. The main area of convection follows suit, which leaves Southeast Asia in the subsiding, dry region. With the eastward shift of the heating, the subtropical jet stream is also shifted eastward. This reduces the distance to the next center over South America and effectively prevents any deflections of the jet winds over the equator

Figure 15 A schematic illustration of the formation of the subtropical jet stream. Air accelerated by the upper-tropospheric PGF gains speed poleward, but is simultaneously affected by the fV, which, by deflecting it clockwise in the Northern Hemisphere (anti-clockwise in the Southern Hemisphere) tries to bring the air back towards the equator. A balance is reached around 30°latitude

pressure is established around 30°latitude. Here we find at 7–13 km height a strong westerly flow with winds of 30–80 m s^{-1}. This is the subtropical jet stream, the largest, strongest and most persistent wind system in the Earth's atmosphere. It also sets the poleward limit of the Hadley circulation (Figure 15).

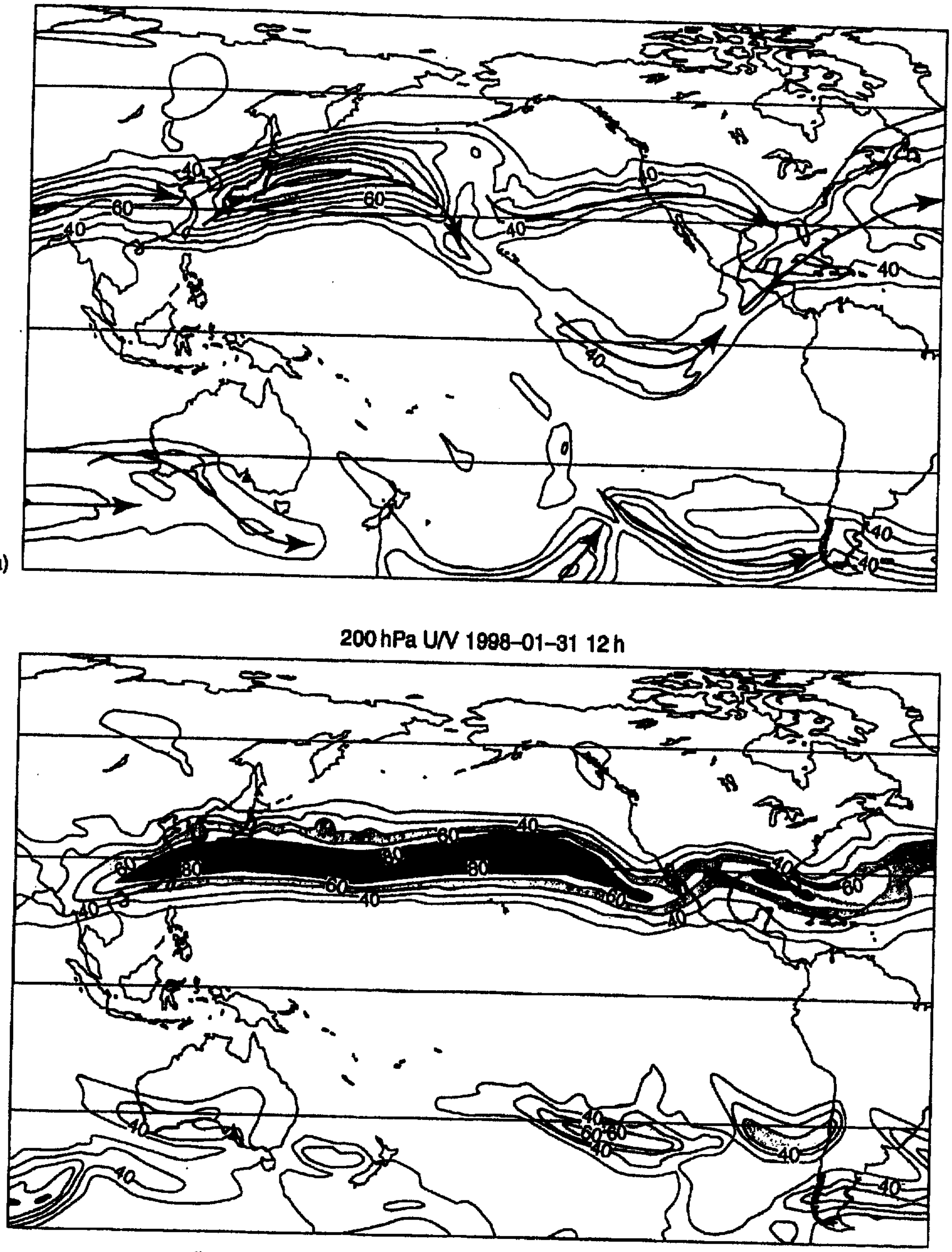
With a faster rotation of the planet, and/or weaker Hadley circulation, the subtropical jet would have been closer to the equator and weaker. If the Earth on the other hand rotated more slowly and/or the Hadley circulation had been stronger, we would have a much stronger jet stream at higher latitudes.

The Longitudinal Variations of the Subtropical Jet

The strongest heating of the atmosphere occurs over the Caribbean and the Amazon basin, over the Sahel and the African rainforests, and over the Indonesian archipelago and Australia. This is where the upper level high-pressure systems are strongest, and it is poleward from these regions where the subtropical jet has wind maxima. Over the oceans, where the heating from below is less intense, the meridional pressure differences are weaker and so are the PGFs.

Due to the irregular distribution of continents and oceans, these regions of maximum heating are not evenly distributed longitudinally. The maximum over East Asia and Australia around 120°E is far upstream of the next downstream maximum over the Americas, which is located around 80°W, almost half the circumference of the Earth. The Subtropical jet stream, which originates over the western Pacific, therefore has a long distance to the next heat source over Latin America. During the long passage over the Pacific, the fV has plenty of time to affect the jet stream, trying to drive it into a circular motion toward the equator. This is facilitated by the SST distribution, with the warmest water in the west Pacific. As the temperature of the ocean surface water decreases eastward, the upper level PGF therefore weakens and allows branches of the subtropical jet occasionally to cross the equator from either hemisphere (Figure 16a).

During an El Niño this does not happen. Then the relatively cold sea surface water in the central and eastern Pacific is replaced by warm water, the heating increases and so has the upper PGF, which prevents the fV from deflecting the winds over the equator. The extra heat source over the central and eastern part of the equatorial Pacific creates an upper-air high-pressure area that effectively prevents the fV from deflecting the wind across the equator (Figure 16b).



(a)
(b)

Figure 16 (a) The winds at 200 hPa (approx. 12 km) on 19 February, 1999. The isolines indicate the wind speed in intervals of 10 m/s. The subtropical jet stream over eastern Asia is almost permanently positioned over southern Japan. Halfway across the North Pacific, due to the action of the fV and the reduced heating from below, the wind can be deflected across the equator. Because the fV deflects winds to the left in the Southern Hemisphere, the jet wind is normally deflected back to the Northern Hemisphere. (b) The same as (a) for 31 January, 1998. This is a typical upper tropospheric flow pattern over the Pacific during an El Niño period with the two subtropical jet streams clearly separated in each hemisphere (*Source:* ECMWF analyses)

Seasonal Variations of the Subtropical Jet

The subtropical jet streams show large seasonal variations, particularly in the Northern Hemisphere. Indeed, there is nothing that shows the close connection between the subtropical jet stream and large-scale heat sources better than the extraordinary change in the subtropical jet that regularly occurs from winter to summer. When summer approaches, the heat source for the atmosphere is extended well into the middle latitudes. Due to strong insolation over the large subtropical land masses of the Northern Hemisphere, the large mass circulation of winter that transports heat poleward is greatly weakened and the westerly subtropical jet stream disappears as a circumpolar phenomenon (Figure 17a,b).

The Caging In of the Atmospheric Machine

The fact that air moving from the equator into the subtropical jet stream is, on the one hand, prevented from moving farther poleward by the fV, and on the other, prevented from returning equatorward by the PGF, causes a congestion of air in the upper troposphere. This is felt at lower levels as a small (1–2%) increase in the pressure. This creates the subtropical high-pressure belts that are found around 30° latitude in each hemisphere underneath

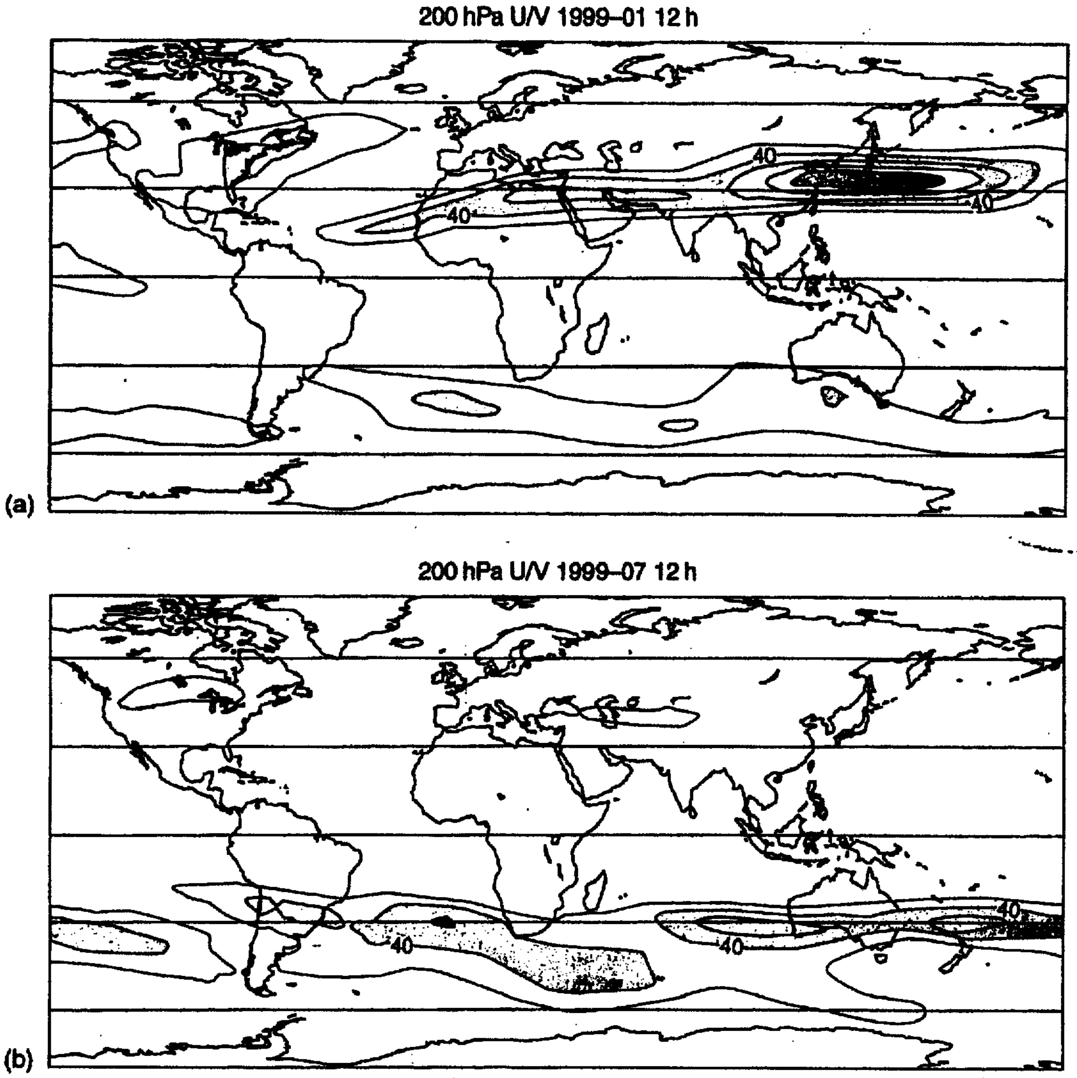


Figure 17 (a) The mean wind speed at 200 hPa (ca 12 km) during February 1999. The Northern Hemisphere subtropical jet stream is clearly seen around 30 °N, whereas its counterpart in the Southern Hemisphere is weak. The mid-latitude jet streams around latitude 50° do not come out strongly on the monthly averages due to their great geographical variability. (b) The same as (a) but for August 1999. The subtropical jet stream in the Northern Hemisphere has disappeared, but its counterpart in the Southern Hemisphere is clearly seen around 30°. The mid-latitude jet streams around 50 °S are still visible due to their strength and interaction with the subtropical jet stream

Figure 18 The subtropical high pressure belt is located beneath the subtropical jet stream. The high pressure is particularly strong over the oceans, which coincides with the left entrance of the jet stream where strong sinking motion occurs

the subtropical jets (Figure 18). The convergence of air at high levels also leads to sinking motions, which in turn leads to a drying of the air and dissipation of the clouds. The clear skies at subtropical latitudes lead to increased outward radiation into space and cooling, which is compensated during daylight hours by dry convection, bringing air heated by the underlying surface to higher levels in the atmosphere.

In the subtropical belt, individual high-pressure centers are formed over the subtropical oceans due to a combination of dynamic and thermal effects. These were called the *Doldrums* or the *Horse latitudes* by seamen due to their weak winds. The subtropical highs are closest to the equator during the respective winters, and displaced poleward during the respective summers.

The Polar Regions – Another Source of Energy

Not only do the tropics supply energy to the atmosphere, but so do the polar regions to a small degree. Around the Pole, air is constantly cooled, which makes it heavier and causes it to sink. As we saw with the ice skater effect, sinking motion is linked with inflow of air at upper levels and outflow at lower levels. The upper air becomes a gigantic inward spiral, the lower an equally large outward spiral. Due to the Coriolis effect and the adjustment between pressure and wind, a low-pressure region is found at upper levels and high pressure at lower levels. From this polar cap cell, air is accelerated away from the polar regions and deflected by the fV to create a regime of northeasterly winds down to 60–70°N. During summertime, with its strong insolation, this process disappears.

WHERE THE ENERGY IS LOST

If it is in the tropics that the Sun's radiative energy is transformed into energy of different kinds, then it is in the mid-latitude westerlies (the roaring forties) that most of this energy, released in storm developments, finally returns to frictional heat and is radiated back into space.

The Mid-latitude Westerlies

The mid-latitude westerlies mirror the tropical easterlies. While the tropical easterlies blow in the opposite direction to the Earth's rotation, and thereby slow it down, the extratropical westerlies speed up the Earth's rotation. These two effects balance, on average, but due to the seasonal variations of the wind regimes, the Earth's rotation, and thereby the length of the day, the Earth's rotation undergoes a periodic variation of the order of milliseconds (*see* **Atmospheric Angular Momentum and Earth Rotation**, Volume 1).

But why do we have westerlies at all in the mid-latitudes? And why are they not, as in the Tropics, compensated at upper levels by winds in the opposite direction? The reason is that the winds in the mid-latitudes are created and maintained by other mechanisms than those decisive in the tropics. One is the effect of the Earth's rotation; while it can be neglected in the tropics, in the mid-latitudes it plays a very important role.

When the wind is accelerated poleward from the subtropical high-pressure belt, it is affected by an increasing fV. A strong adjustment between wind and pressure is continuously taking place. As a result isobars typically become oriented from WSW to ENE in the Northern Hemisphere, WNW to ESE in the Southern Hemisphere. Where the fV tries to drive the air equatorward, back towards higher pressure, the pressure gradient sharpens. Part of the kinetic energy of the winds is then converted back into potential energy, which delays its further transport up into the mid-latitudes. Pressure decreases with height more slowly in warm air than in cold, so the meridional temperature contrast gives rise to pressure differences at upper levels. Due to the strong fV, the wind and the pressure field rapidly adjust, resulting in wind from a westerly direction.

The strong fV is also the reason why, on average, the subtropical air masses are confined to latitudes equatorward of 45°. Also, at high latitudes the strong Coriolis deflection makes it difficult for the cold air masses from the high-pressure areas around the poles to extend very far. Arctic air masses are normally found poleward of 65°latitude,

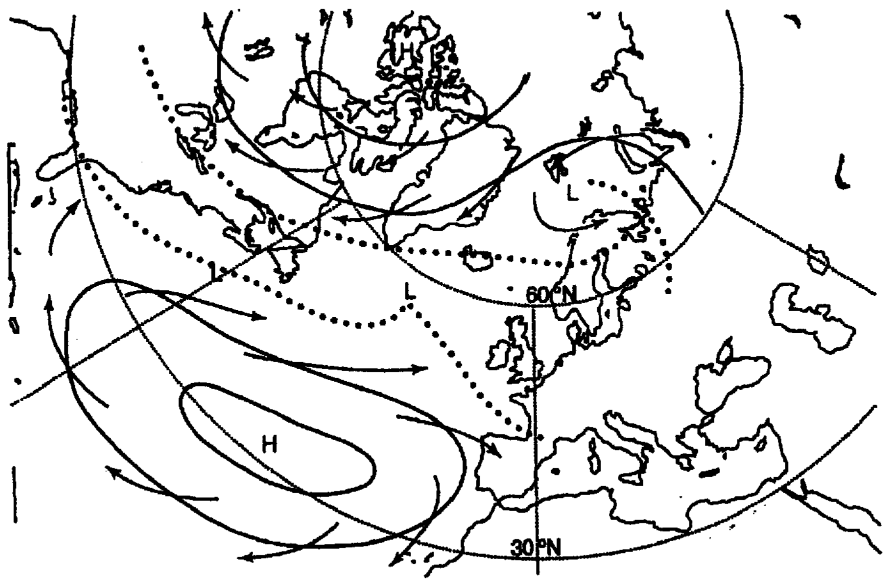


Figure 19 The acceleration of the winds out from the subtropical high pressure belt and the polar cap high pressure system. Winds moving poleward from the subtropical high pressure belt encounter an increasing fV, and a strong adjustment with the pressure field takes place. The fV acts to confine air masses so that cold air from the poles cannot easily penetrate equatorward of 60–65° and subtropical air masses not poleward of 40–50°. The air mass boundary around 45° is called the Polar Front, which in winter extends from the southwestern USA across the Atlantic to the British Isles, and from the northern Mediterranean eastward into Asia. The borderline around 60°, the Arctic Front, is most pronounced over Alaska, western Canada and northen Europe. Another branch of the Polar Front extends from the waters south of Japan across the Pacific Ocean to the US-Canadian border

and then only in wintertime. In summertime the insolation prevents cold air masses from forming. The intermediate latitude band between 45° and 65° is occupied by stagnant air, either from the tropics which has cooled, or from the Arctic regions, which has warmed (Figure 19).

Jet Streams and Fronts

In contrast to the tropics, where the main motion is due to vertical temperature contrasts that create convection that heats the atmosphere, the important driving forces in the mid-latitudes are horizontal temperature contrasts between air masses of different temperature and density. Through the action of the fV, these contrasts are concentrated in narrow zones, so called fronts or Polar Fronts.

In the broad river of air which constitutes the mid-latitude westerlies, long, narrow bands are imbedded between 8 and 13 km altitude where the winds reach 30–80 $m\,s^{-1}$, and are on rare occasions 100–150 $m\,s^{-1}$. These are the Polar Front jet streams. In contrast to the subtropical jet, which gains its strength from winds accelerated in the Hadley circulation, the Polar Front jet streams thrive on the thermal contrasts along the Polar Front (*see* **Fronts**, Volume 1; **Jet Stream**, Volume 1).

The fronts and their associated jet streams do not form a homogenous, uninterrupted band around the hemisphere, but rather separate bands stretching over 60–90° longitudinal sectors where the temperature contrasts and the horizontal pressure differences are more concentrated and the winds stronger than in the intermediate sectors.

As the air moves through the non-uniform pressure field, it is constantly accelerated at the upstream end of a major jet stream, decelerated at the downstream end. With the acceleration, kinetic energy is drawn from the potential energy associated with the pressure and thermal contrasts; with the deceleration, kinetic energy is converted back to potential energy associated with the pressure and thermal contrasts. When the potential energy decreases upstream, these contrasts (gradients) weaken; when potential energy is increased downstream, the contrasts are strengthened. This has the consequence that the whole pressure and temperature pattern will slowly be displaced downstream, while the wind rapidly moves through (Figure 20a–c).

Vertical Motions around Fronts and Jet Streams

Cold and warm air masses of different densities would under buoyancy forces (gravity) tend to arrange themselves horizontally, with cold dense air sinking under the warm light air. Doing so, the system's center of gravity is lowered. But this re-arrangement would involve horizontal motion that would immediately be affected by the fV, which would try to bring the air back. The frontal surfaces are left to slope when a balance is reached between buoyancy forces, which try to create a horizontal stratification, and

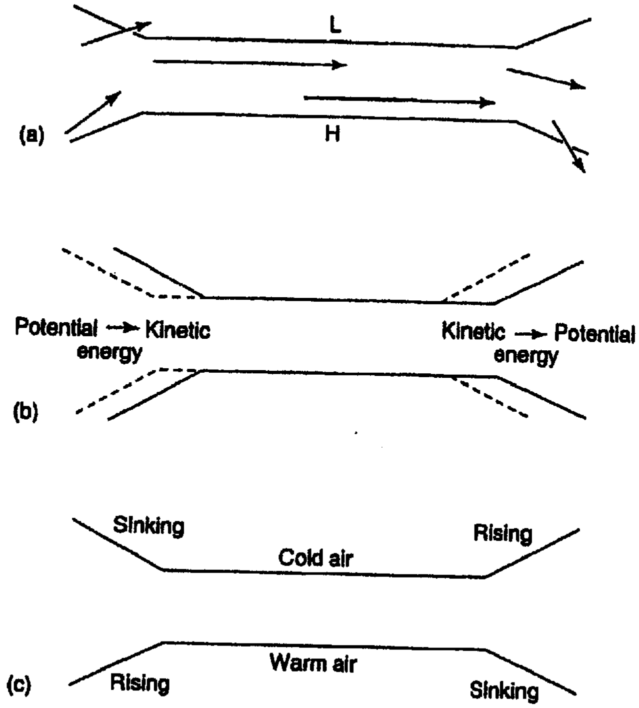


Figure 20 (a) The inflow and outflow in a jet stream. The acceleration of the air at the entrance implies an increase of kinetic energy at the expense of potential energy, the deceleration at the exit implies a decrease of kinetic energy and an increase of the potential energy. (b) Because the level of potential energy is associated with the strength of the pressure differences, these decrease at the entrance and increase at the exit, which leads to the whole pressure system being slowly moved in the direction of the flow. (c) A decrease in potential energy means that the system's center of gravity is lowered. This implies rising of warm, light air and sinking of cold heavy air. With warm air on the equatorward side and cold air on the poleward side, this means at the entrance of the jet rising motion to the right, sinking motion to the left. With a similar reasoning, it can be understood why, at the exit of the jet where potential energy is increasing, cold air to the left is rising and warm air to the right is sinking

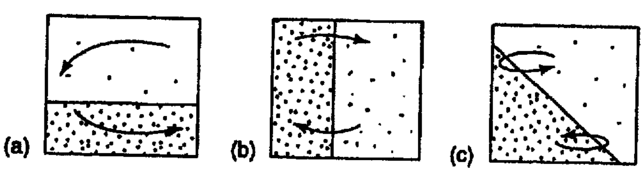


Figure 21 The motion at frontal boundaries is determined by two opposing mechanisms: (a) the thermal contrast tries to create a *horizontal* stratification; (b) the rotation of the Earth tries to create a *vertical* stratification. (c) Neither of the mechanisms is dominant, leaving the frontal surfaces to slope with an inclination of the order of 1:50 to 1:200

the rotation of the Earth, which tries to create a vertical stratification (Figure 21a–c).

The reversal of this process, against the buoyancy forces, and thus against gravity, involves lifting of cold air and sinking of warm air. Because this would raise the center of gravity and increase the potential energy, it can only occur through external mechanical forcing. It is the fV that provides this mechanical forcing when it, being stronger than the PGF, drives the air towards higher pressure, which slows down the wind. Note that it is the PGF, not the fV, that slows down the wind. This can be compared to a ball rolling down a hill. Gravity accelerates the ball when the kinetic energy, increases. Inertia might then drive the ball uphill, but the work, the conversion from kinetic to potential energy, is still made by gravity slowing down the speed of the ball.

The induced vertical motions affect the temperature, sharpening or even creating fronts. This conversion between potential and kinetic energy is reflected in the waves that develop and the accompanying changes of the frontal slopes: flattening when kinetic energy is created, steepening when the kinetic energy returns to potential (Figure 22). During these transitions the frontal surface oscillates like a membrane. Frequently, for reasons which are not well understood, the waves along the frontal boundary escalate into intense vortices.

The Development of Extra-tropical Storms

The subtropical highs are created and maintained by air at higher levels converging into the subtropical jet stream.

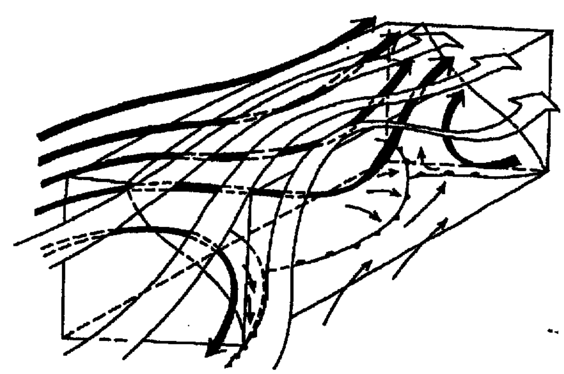


Figure 22 The circulation in extra-tropical weather systems involves a complex interplay of rising and descending motions. Air masses coming from higher and lower latitudes converge in the western part, the "entrance," and diverge in the eastern part, the "exit". While staying separate, they undergo large vertical displacements. Whereas at the left entrance and right exit, where the air is sinking, there is a tendency for downward motion and dynamical stability, at the right entrance and left exit of a jet stream, where the air is rising, there are favorable areas for storm development. The combined sinking of cold air and ascent of warm air represents a conversion of potential to kinetic energy, which maintains the jet stream. Heat and moisture is transported downstream, maintaining and even sharpening and extending the frontal zone

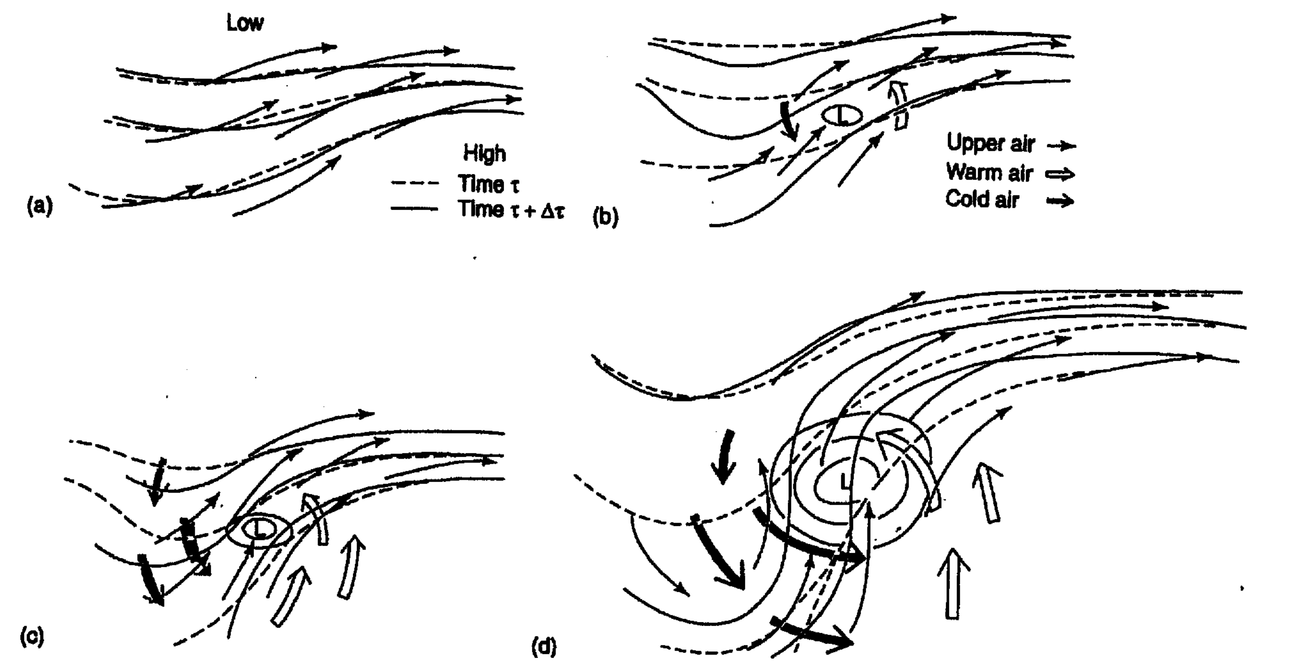


Figure 23 (a) The initial development of a deepening storm. The zone between warm air to the south and cold air to the north; strong pressure differences and a jet stream have developed at high levels. The wind is accelerated into the jet stream and the gradient is weakening [(- - - - -) lines for equal pressure at an early stage, (——) at a later stage]. (b) More air is exported out of the region than imported. This upper divergence causes the pressure to fall at lower levels. A new circulation is rapidly established due to the balance between the PGF and the fV. (c) The transport of warm air northwards makes the upper flow more anticyclonic and increases the acceleration of air out from the region; the transport of cold air southward makes the flow more cyclonic and weakens the acceleration of the wind. (d) The changes in the pressure field lead to an increase in the wind divergence. A chain reaction has come under way with the pressure at lower levels falling, increasing the thermal transports and the amplification of the flow pattern

The opposite process, divergence of the wind at higher levels, creates pressure drops at low levels that often initiate the development of mid-latitude cyclones. There are various mechanisms causing upper-air divergence, the most common of which is when the wind is accelerated into or out of a jet stream. The wind is subject to a strong acceleration, partly reflected in the curvature of the wind flow, from a cyclonically (anti-clockwise) curved flow to an anticyclonically (clockwise) one in the Northern Hemisphere (Figure 23a).

If the air leaving a region is not replaced by the same amount of air coming in, this is felt at lower levels as a drop in pressure. As the wind at lower levels adjusts to the pressure field, a cyclonic circulation is created (Figure 23b). This circulation affects the temperature distribution, which close to jet streams is normally characterized by strong horizontal temperature differences.

East of the low-pressure system the wind at lower levels transports warm air poleward. Pressure decreases more slowly upward in warm air and the influx of warm air increases the pressure at upper levels. This intensifies the anticyclonic circulation at upper levels and accelerates the wind even more. As a consequence, more air is exported out of the region (Figure 23c).

To the west of the low, the induced wind circulation transports cold air equatorwards. Because pressure decreases more rapidly upward in cold air, the influx of cold air decreases the pressure at higher levels and intensifies the cyclonic (anticlockwise) circulation. The increased curvature of the airflow further slows down the wind velocity and less air is imported into the region.

With more air exported, and less air imported, the pressure at lower levels will continue to fall, the circulation around the low-level system will intensify, and so will the transport of warm air northward and cold air southward (Figure 23d). A chain reaction has started that leads to a major storm development with strong winds circulating around the low pressure area, warm air rising above cold air at the front of the system, and cold air sinking below warm air at the rear of the system. It comes to an end when the circulation has spread from lower levels to higher levels and established the warm air in a position above the cold air.

Upper air divergence can also be caused or enhanced by latent heat release when moist air condenses. The heating of the air raises the pressure at higher levels and further enhances the upper outflow of air. This is the main driving mechanism in tropical and subtropical storms, but also plays an important role in extra-tropical developments. The latent heat release at the right entrance of the jet maintains the thermal contrasts in the upstream part of the jet and keeps it in a westerly position where new cyclones can form. This can lead to a succession of frontal cyclones forming and, within a couple of days, to the creation of a cyclone family.

Downstream Development

The kinetic energy released during the cyclonic development takes two routes: one into the storm itself, where it is soon to a large extent lost in the frictional process at the Earth's surface; the other route into the mid- and upper-troposphere jet winds where it is rapidly transported downstream. Arriving in the next system downstream, these huge amounts of energy can have a profound effect, either directly or through conversion back to potential energy through the action of the fV (forcing the PGF to do negative work on the air). Although the density of the air in the upper troposphere is only 1/3 of the density at the ground, the wind velocities are about ten times stronger, which makes the kinetic energy 30 times larger. The kinetic energy of an upper level jet stream of $60\,m\,s^{-1}$ is the same as a hurricane force wind at the ground of $35\,m\,s^{-1}$.

When the downstream storm develops, which receives the energy in the same way, it transports part of its released kinetic energy into the next system downstream, and so on. As long as the dynamic conditions are favorable, the influence of one storm rapidly spreads downstream to the next in a hemispheric domino effect (Figure 24). The speed of this "downstream development" is roughly $30\,m\,s^{-1}$, which corresponds to $30^\circ\,day^{-1}$ at 45°latitude, but slightly less in summertime. This should be compared with the typical phase speed of $10^\circ\,day^{-1}$ of a normal frontal cyclone wave. In the Southern Hemisphere the typical velocity of the downstream influence is $40^\circ\,day^{-1}$ due to the predominant zonal flow.

The fact that the influence of a weather development spreads three times faster than the weather system has consequences for weather forecasting. A five day weather forecast for Europe is dependent on the initial atmospheric conditions over North America, a five day forecast for North America on the conditions over the Northwest Pacific, although the weather systems in those areas may not necessarily arrive in the target area.

Cut-off Lows and Blocking Highs

The downstream development not only deepens cyclones, it also amplifies high-pressure systems. A strong explosive cyclogenesis creates strong pressure differences high up in the troposphere that force the jet stream poleward and helps to amplify a downstream ridge. The poleward transport of warm air further increases the pressure and supports the creation of a high pressure system through a deep vertical layer, a so called *blocking high*. The upper current moving over the ridge, partly due to the anticyclonic curvature, accelerates and make the fV stronger than the PGF. The wind, thus driven towards higher pressure by the fV, creates a low-pressure system, a so-called cut-off low (Figure 25).

Flow Patterns Forced by the Earth's Surface

The uneven heating and cooling of the oceans and the continents also contribute towards shaping the large-scale

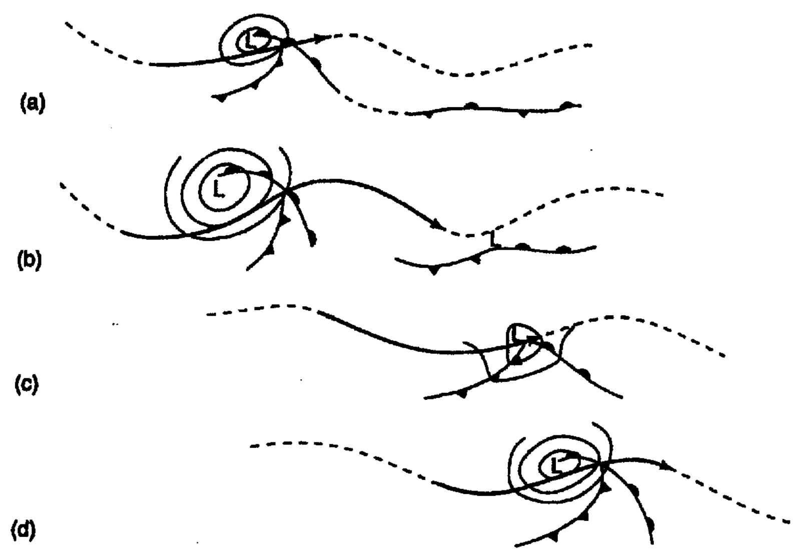


Figure 24 Down stream development. The kinetic energy released during the development of one storm (a) is transported downstream in the jet stream (b) and in a short period of time affects the next system (c),which then during its development releases additional kinetic energy into the upper tropospheric flow (d)

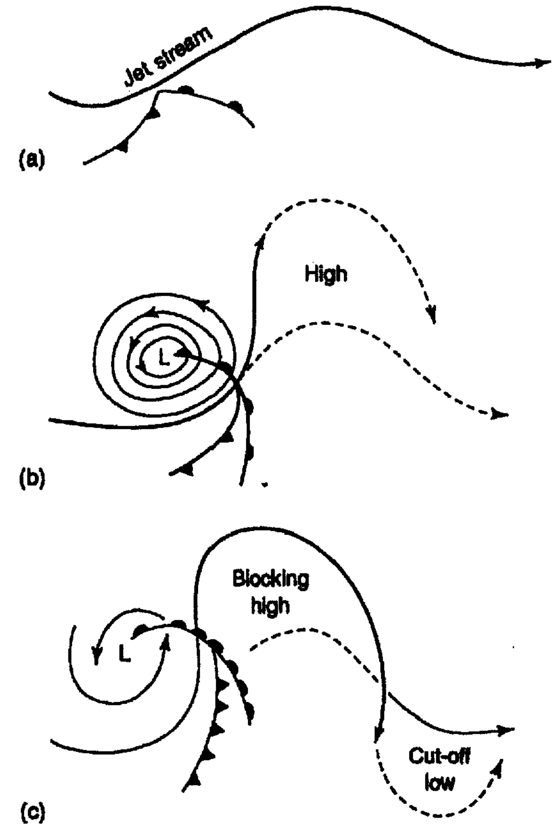


Figure 25 (a) Blocking highs tend to develop as a response to explosive cyclones. (b) When huge amounts of kinetic energy are ejected into the upper atmospheric flow, bringing into a new direction and often amplifying a downstream high pressure system. (c) When the strong winds in the free atmosphere reach the downstream side of these amplified high pressure systems, there is a tendency to form low pressure circulations, so called "cut-off lows"

flow over the Earth. In wintertime there is a strong cooling over the continents due to radiative losses. The cooled air has a general tendency to sink. This sinking motion is associated with upper inflow and, in line with the ice skater effect, cyclonic circulation. At lower levels there is outflow and anticyclonic circulation.

When the cold air is transported out over surrounding warm ocean waters, which normally are to the east of the continents, there is widespread and intense rising motion due to heating from below. This creates favorable conditions for low pressure and low level cyclonic circulation. Cyclones develop easily and strongly by drawing from these thermal contrasts, which are found in the western Atlantic and the northwest Pacific. In summertime the heating of the continents contributes to rising motion with low level inflow and upper level outflow, with a tendency for anticyclonic circulation.

Another influence, which contributes to the shape of the large-scale flow, is from the major mountain ranges. The atmospheric flow tends to accumulate air on the windward side, creating or strengthening a PGF across the mountain. This accelerates the wind to an extent that the fV becomes stronger than the PGF on the lee-side of the mountain ridge. The wind therefore starts to deviate towards higher pressure and, similar to the cut-off process described above, creates a cyclonic circulation and a low-pressure system. The tendency to have cyclonic shaped flow in the lee of major mountain ranges facilitates the development of cyclones in this area, whereas there is a enhanced tendency for high pressure systems on the windward side.

The waves generated by a combination of land-sea contrasts and the orography, further enhanced by cyclone developments where the dynamic conditions are favorable, ultimately create geographically very large waves or weather zones, so called Planetary or Rossby waves (*see* **Rossby Waves**, Volume 1). They are responsible for periods of persistent dry or wet, cold or warm weather types. The so called *beta-effect*, the latitudinal variation of the fV, is particularly prominent in systems of large north–south extent and may make the waves become stationary and even move slowly westward, against the general current (*see* **Coriolis Effect**, Volume 1).

A SUMMARY OF THE ATMOSPHERIC CIRCULATION

A simple two-dimensional summary of the general circulation of the atmosphere must be accompanied by some words of caution when the picture is interpreted. As was discussed in the Introduction, average values do not necessarily represent what is typical and it is not normally possible to deduce the path of air particles from average wind fields.

Two Ways to Look at the Meridional Circulation

The commonly depicted, tri-cellular, cross-section of the atmosphere with rising motion in the tropics and the mid-latitudes, sinking motion in the polar regions and subtropics should not be interpreted as showing the approximate path of any individual air parcel. The cross-section is based on averages of winds in fixed locations (called Eulerian averages) (Figure 26a). A picture of the average routes of air parcels can only be accomplished by averaging in terms of trajectories (called Lagrangian averaging). If one wishes to determine the likely spread of a pollutant in the atmospheric circulation, the Lagrangian view would give a much better idea than the Eulerian (Figure 26b).

The Eulerian and Lagrangian methods do not necessarily have to show different pictures. Due to the steadiness of the Hadley cell, the two averages portray similar pictures of the rising branch within the summer hemisphere and large mass transports into the winter hemisphere. But at subtropical and middle latitudes, where the motion is more complicated (see

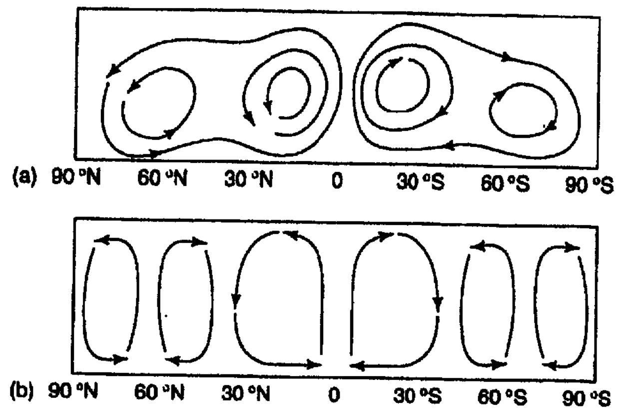


Figure 26 Cross sections of the atmospheric circulation using: (a) Lagrangian averages; (b) Eulerian averages. A common mistake is to visually interpret the Eurlerian average, which denotes streamlines, as if it were Lagrangian, which represents the tracks of individual air parcels (trajectories)

Figure 11) the two averages display very different images. Instead of three cells in the Eulerian picture, the Lagrangian view depicts one single cell extending from the tropics to polar latitudes.

Correlated Fields

The (Eulerian) average wind flow depicts rising motion at mid-latitudes between 50° and 70°, sinking at subtropical latitudes around 35°. Because the air is colder at higher latitudes than lower, such a circulation seems to imply rising of cold air and sinking of warm, which would imply an increase in potential energy. This interpretation has supported the notion that between the Polar Cap cell and the tropical Hadley cell, both sources of kinetic energy, there is a third cell, the *Ferrel cell*, that is supposed to convert kinetic energy back to potential. But this notion of a mid-latitude cell is due to a misinterpretation of statistics.

It is true that the 50°–70°latitude band has a tendency for upward motion, but these are mostly associated with rising tongues of warm air in connection with passing cyclones. The same applies for the subsidence around 35° subtropical latitude, which is associated with outbreaks of cold air that are generally associated with sinking motion (Figure 27). The relevant correlation, the one which can be given a physical interpretation, is not between mean values, but between instantaneous values of temperature and vertical velocity.

The Atmosphere's Energy Budget

The interpretation of the atmosphere's energy budget must be done with care. On a long-term average, the potential energy of the atmosphere is converted into kinetic, which, in turn, is lost in friction against the Earth's surface and internal friction (viscosity) (Figure 28a). On average, the contributions from the Sun to the potential energy makes up for the average loss in friction. However, this is the

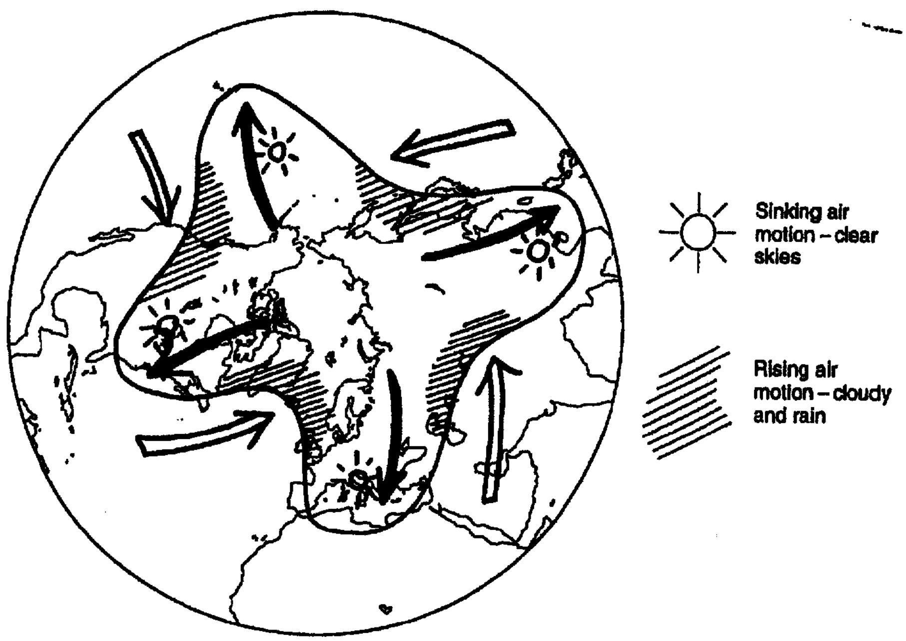


Figure 27 Advances of warm air poleward is often associated with rising motion (denoted by hatching symbolizing precipitation) whereas cold outbreak to low latitudes often are accompanied with sinking air and rather cloudless conditions (here symbolized with a sun)

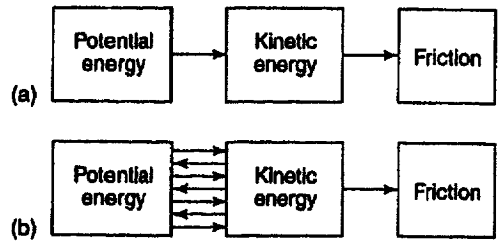


Figure 28 (a) The net energy budget shows that potential energy is lost to kinetic energy, which in its turn is lost to friction. (b) Only the kinetic energy to friction conversion is irreversible, the potential to kinetic is highly reversible

net effect, or budget, and does not necessarily tell what is going on at every moment at any place where there is a constant conversion between potential and kinetic energy (Figure 28b).

In the atmosphere, it is the subtropical high pressure belts that act as sources for kinetic energy; the low pressure systems in the mid-latitudes act as sinks for kinetic energy due to frictional losses. But before the kinetic energy is lost to friction, it has been "recycled" several times through reversible conversions between potential to kinetic energy in the mid-latitude cyclones, most clearly reflected in the downstream development process.

A schematic cross-section of the atmosphere is given in Figure 29. The Hadley cell circulation should be seen as a longitudinally elongated spiral into the diagram where air rising at one longitude descends at another far downstream. Due to the seasonal variations of the maximal heating from the Sun, the Hadley cell is slowly oscillating around the equator and also varies in intensity.

The fronts and jet streams in the westerlies are drawn in a way to suggest their general orientation from WSW to ENE. The vertical motion in the area might on average be upward, but, again, this is just the net result of very strong up- and downward motions.

The Stratosphere

The flow in the lower stratosphere, the part of the atmosphere at 10–25 km height, is generally characterized by a rising motion of cold air from the equatorial belt in the summer hemisphere to a sinking motion of warmer air in the mid-latitudes and around the poles. With this circulation, converting kinetic energy into potential energy, its energy cycle differs considerably from that of the atmosphere as a whole. It is rather like a heat engine in reverse, or a thermodynamic refrigerator. This can only come about through mechanical forcing, and it is the motions in the atmosphere below that provide this mechanism (*see* **Stratosphere, Temperature and Circulation**, Volume 1).

However, the stratosphere also has dynamics of its own where instabilities create strong vortices or blocking highs, so called *sudden warming*. The dynamics of the stratosphere is further complicated by a complex interaction between radiative, dynamical and photochemical processes in the ozone layer.

The stratosphere has been under intense study since it was discovered 100 years ago, still new discoveries

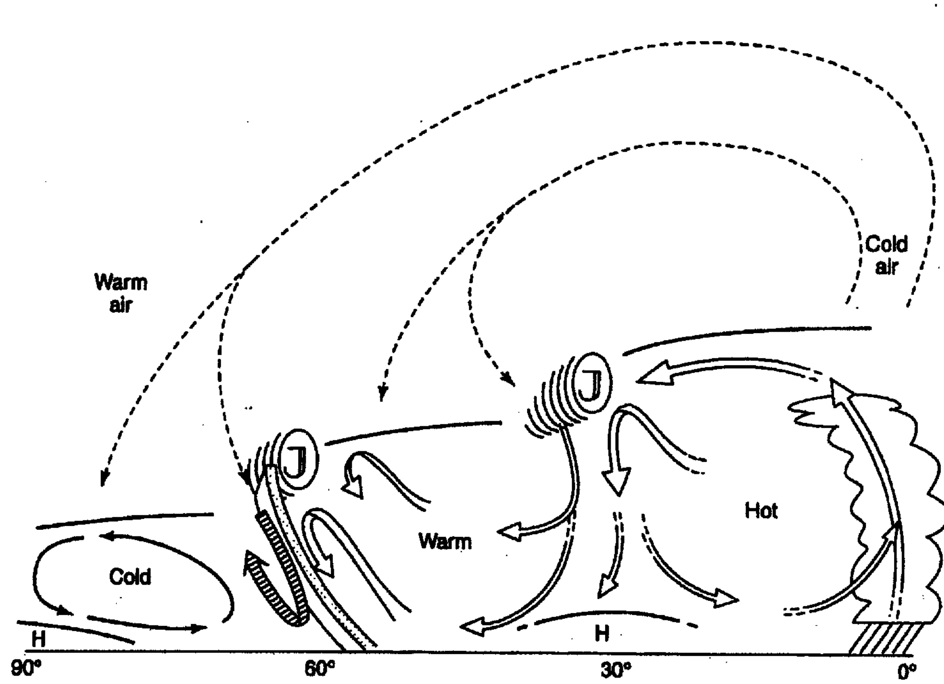


Figure 29 A schematic cross section of the atmosphere (troposphere and lower stratosphere)

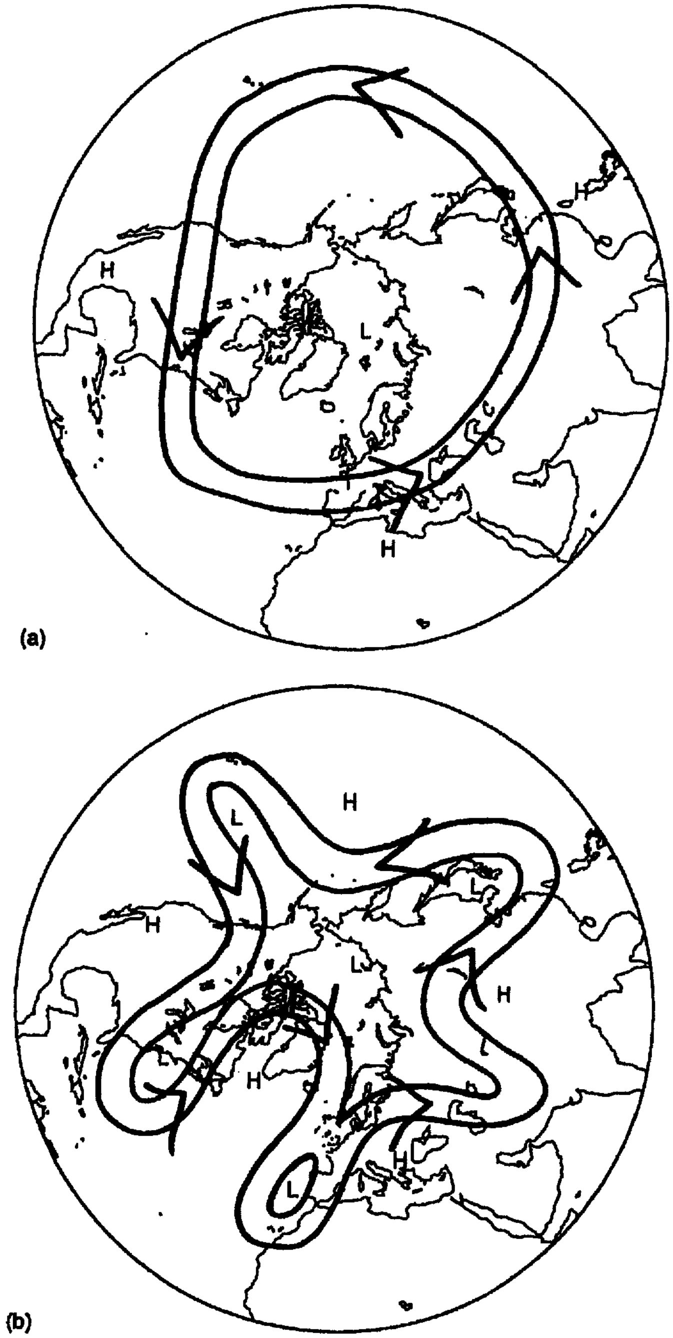


Figure 30 The atmospheric flow vacillates between two regimes, (a) one zonal (mostly east-west flow) and (b) one meridional (much north-south flow) with pronounced high pressure formations ("blocking") and small, isolated low pressure systems ("cut-offs"), both created when there are strong imbalances between the pressure field and the winds

are made almost every decade. So for example, a 26-month oscillation in the upper winds has been found in the tropics. This Quasi-Biennial Oscillation (QBO) is still not fully understood, nor the general interaction between the stratosphere and the underlying atmosphere (*see* **Quasi-Decadal Oscillation**, Volume 1). There are still discoveries to be made and surely many surprises in store!

Can the Atmospheric Flow Change?

There has been much research into the dynamics of the large-scale flow to acquire a deeper understanding of past climate changes (the ice ages) and possible scenarios for the future. It is not yet clear why the atmospheric flow can be anomalous for relatively short periods like several months or years. Why were the mid-1990s characterized by dry conditions over Western Europe, why have the last five years been wetter than before?

One process where an understanding is growing concerns the mechanisms that induce wave patterns because the land-sea contrasts and the orography may not necessarily pull the atmosphere in the same direction. There is also a seasonal variation in the land-sea contrast, which is only slightly evident in the orographic forcing. Due to this misfit, the large scale flow is never quite in a stable state and tends to vacillate between two preferred states: one relatively straight zonal type with a well-developed and fairly broad westerly current in middle latitudes, and the other large meridional flow patterns which extend meridionally with strong cyclonic vortices at low latitudes and anticyclonic flow at high latitudes (Figure 30). Each pattern persists for several weeks, whereas the transition from one to the other is relatively brief.

There are reasons to assume that the typical flow regimes that we regard as normal might be just one of several possible quasi-stable states of the atmosphere. Any change of the overall temperature of the atmosphere will most likely change the flow pattern as well. This might have unexpected consequences; an over-all warming of the atmosphere might lead to some regions getting colder temperatures due to the change of flow regimes. This can only be fully understood if the motion of the general circulation also includes the motions of ocean currents. It might be that neither the atmosphere nor the oceans can undergo any radical changes on their own, they might only be able to change if they change together.

Atmospheric Processes and Interactions

see Earth System Processes (*Opening essay*, **Volume 1**)

Atmospheric Radiation Measurement Program (ARM)

see ARM (Atmospheric Radiation Measurement) Program **(Volume 1)**

Atmospheric Structure

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Atmospheric structure is usually represented in terms of temperature profiles which define the boundaries of the troposphere and stratosphere. It is also possible to describe the structure of the atmosphere in terms of composition. The best known example of this is the ionosphere where free electrons abound. However, high in the atmosphere gases separate gravitationally (heterosphere) and at its extreme edges (exosphere) molecules so rarely collide that they can undergo ballistic trajectories and escape from the atmosphere.

The atmosphere appears to be a transparent gas and initially devoid of obvious structure. However, more careful examination reveals that there are layers. Clouds, especially those with flat bases make this fairly obvious.

Although there is a gradual decrease in pressure with altitude, it is the thermal changes with altitude that are most often used to define structure. Through the lowest part of the atmosphere, the troposphere, temperature decreases with altitude at about 6.5 °C km^{-1} (*see* **Lapse Rate**, Volume 1). Air closest to the ground is warmest and this is the region we are most familiar with. It is here that our weather takes place. Convection can be active in the troposphere and is dramatically seen in the formation of thunderclouds. At times, air close to the ground can become cold, particularly at night. Under these *inversion* conditions, the air is resistant to vertical mixing, pollutants can accumulate or fogs can form.

Higher up in the atmosphere the absorption of radiation by the ozone layer warms the atmosphere. Thus, temperature increases above the *troposphere* (Figure 1), in a layer called the *stratosphere*. The increase in average temperature with height means that the stratosphere is very stable with respect to vertical mixing, hence the tendency for well defined strata of air to form. Yet it is so cold that water is present in very small amounts and clouds are very rare and

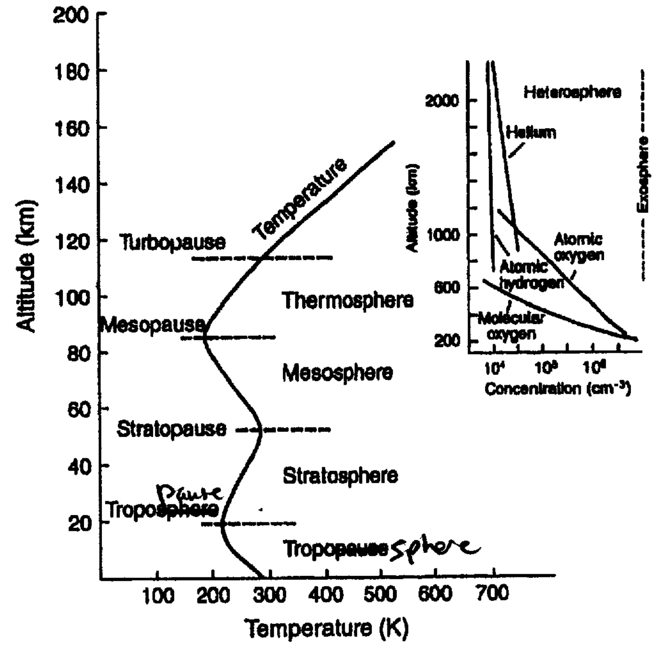

Figure 1 The structure of the atmosphere. Note how the boundaries at the top of each of these layers are called the tropopause, stratopause, mesopause. (Adapted from Andrews *et al.*, 1996)

consist of sulfuric and nitric acid containing droplets or icy crystals.

Above the stratosphere, the air cools again in the *mesosphere*. Here, noctilucent clouds can be observed in the summer at high latitudes. These are formed through water nucleation onto dust or ions, possibly from meteor trails, which penetrate the mesosphere. The water probably derives from the photooxidation of methane, which has led to speculation that increases in atmospheric methane has caused an increase in the frequency of noctilucent clouds.

Even higher, we encounter the *thermosphere*, where the atmosphere has a very low density and the molecules are warmed by direct impact of hard photons from the Sun. Its high temperature is sensitive to output of solar radiation, which varies as the Sun passes through its quiet and active phases every 11 years. Although the temperatures are very high, the gas is so tenuous that it carries little heat, so does not control the temperature of objects such as spacecraft operating at these altitudes.

There are other ways we can define atmospheric structure. The lower atmosphere is well mixed, where turbulent processes are stronger than the gravitational settling of the gases. The top of this well mixed part of the atmosphere, above which gases settle out dependent on their molecular weight, is called the *turbopause*. Below the turbopause the atmosphere is often termed the *homosphere* where the gases are well mixed. Above the turbopause it is called the *heterosphere* where we find the light gases beginning to dominate at the top of the atmosphere (Figure 1).

The inset of Figure 1 shows the heterosphere and we can see that the gases have separated out due to gravitational settling, but many are not the familiar gases we know at ground level (*see* **Atmospheric Composition, Present**, Volume 1). Even oxygen and nitrogen are not in their familiar diatomic form. The photodissociative processes at high altitude break the molecules down into atoms, thus, at the very top of the atmosphere we find a dominance of hydrogen and helium.

These divisions of the atmosphere have only considered the neutral molecules. The upper atmosphere is affected by ions in a region known as the *ionosphere*. These ions are at relatively low concentration, but their effects are so pronounced that they have been much studied. In particular the way they affect radio waves has been of importance for much of the 20th century. The ionosphere is structured into a number of regions. Early studies used radio waves as probes, which detect changes in electron density in the upper atmosphere. This meant that early studies focussed on the electron distribution of the ionosphere, although scientists were well aware that the charge would be balanced by positive ions. The advent of rockets allowed the positively charged species to be identified.

In the upper atmosphere incoming meteors, although contributing only small amounts of material, have noticeable effects because the atmosphere is so tenuous. As the meteors ablate volatile materials such as water they leave a trail. In addition, less volatile materials such as sodium, magnesium and iron are ablated and form distinctive layers.

At the very outer extent of the atmosphere we encounter the *exosphere*. Here, the gas is so tenuous that molecules follow ballistic trajectories, kilometres in length before colliding with another molecule. Some of the molecules travel so fast that their ballistic trajectories at thermospheric temperatures allow them to escape from the atmosphere completely and become lost to space. In the case of the upper atmosphere of the Earth it is hydrogen and to a lesser extent helium that escape with residence times of about a thousand and many millions of years, respectively. The rapid escape of hydrogen is important as it allows atmospheres to become more oxidizing over time.

See also: **Atmospheric Motions**, Volume 1.

REFERENCE

Andrews, J E, Brimblecombe, P, Jickells, T D, and Liss, P S (1996) *An Introduction to Environmental Chemistry*, Blackwell Science, Oxford.

FURTHER READING

Brekke, A (1997) *Physics of the Upper Polar Atmosphere*, John Wiley & Sons, Chichester.

H

Hadley Circulation

The *Hadley circulation* is one of the simplest and best-known conceptual pictures (models) of the prevailing circulation of the Earth's atmosphere (*see* **Atmospheric Motions**, Volume 1). George Hadley originally described the conceptual model in 1735, based on a combination of sparse, mostly near-surface, atmospheric observations and 18th century physical understanding. Despite these limitations, the conceptual model is still valid, with some modifications.

In the modern picture, the stronger solar heating at low latitudes relative to high latitudes leads to rising air at or near the equator, and sinking air near the poles. To close the circulation, air flows towards the poles in the upper levels of the atmosphere, and towards the equator near the surface. There are then two equator-to-pole circulation patterns, or Hadley cells, one in each hemisphere. The Hadley cells convert potential plus thermal energy to kinetic energy. As a by-product of this conversion, they transport heat from low to high latitudes, cooling the lower latitudes and warming the higher latitudes.

The Hadley circulation also exerts a profound influence on the prevailing atmospheric winds, producing the eastward jet streams in the mid-latitude upper troposphere, and the low-latitude surface easterlies and mid-latitude surface westerlies. Because the Earth is spinning eastward on its axis, and the distance of the atmosphere from the axis of rotation decreases from equator to pole, the poleward moving air in the upper troposphere acquires eastward velocity relative to the ground, due to the tendency to conserve angular momentum (much like a spinning skater pulling in his or her arms). Similarly, air in the equatorward moving lower branches of the Hadley cells eventually acquires westward velocity, but with weaker magnitude than the upper level winds due to surface friction. As the field of meteorology developed after Hadley, an important modification to Hadley's original picture was to take into account the passage of air through storm systems at middle and high latitudes, which reduces the eastward velocities in the upper levels, and causes apparently poleward prevailing midlatitude surface winds.

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Halocarbons

Halocarbons are a category of chemicals containing carbon and at least one of the halogens; fluorine, chlorine, bromine and iodine. Compounds in this category are of both natural and anthropogenic origin. Applications for the human-produced halocarbons include pharmaceuticals, crop protection, refrigerants, plastics as well as a wide range of other industrial and consumer products. Halocarbons involved in global environmental change possess three common characteristics: the compounds are persistent in the environment (generally decades to centuries), mobile in the environment, and either unique to the environment or they are emitted in sufficient quantities to provide a significant enhancement to a natural background concentration. Several classes of gaseous halocarbons that contain chlorine or bromine, including chlorofluorocarbons (CFCs) (*see* **Chlorofluorocarbons (CFCs)**, Volume 1), hydrochlorofluorocarbons (HCFCs), halons and some chlorocarbons, are involved in the stratospheric ozone-depletion issue. Production and use of these compounds are controlled under an international agreement, the Montreal Protocol on Substances that Deplete the Ozone Layer (*see* **Ozone Layer: Vienna Convention and the Montreal Protocol**, Volume 4). Those classes of compounds plus other gaseous halocarbons including hydrofluorocarbons (HFCs) and perfluorocarbons (PFCs) are involved in the global climate change issue. HFCs and PFCs are among the compounds listed for control under the Kyoto Protocol to the United Nations Framework Convention on Climate Change. Several of the higher molecular weight chlorocarbons are implicated in a third global environmental change issue, that of persistent organic pollutants (POPs). In addition to the three

J

Jet Stream

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The jet stream or band of maximum winds in the upper troposphere was first recognized as an important global phenomenon in 1944. Normally it consists of two segments in the Northern Hemisphere, one from western North America to Northern Europe, and the other from western North Africa across Asia and the Pacific. Such segments tend to spiral toward the pole in both hemispheres, and change latitude with the seasons. The main jet streams are in the westerlies, but there are also some easterly jet streams, mostly in the tropics.

DEFINITION

The standard glossary definition of the jet stream is:

> Relatively narrow river of very strong horizontal winds (usually 50 knots or greater) embedded in the planetary winds aloft. These jets are typically located in the upper troposphere above regions of strong horizontal temperature contrasts (fronts) .... Several major jet streams include the polar jet and the subtropical jet.
>
> (Geer *et al.*, 1996)

The definition omits the fact that these elongated west–east bands are nearly always composed of high-speed westerlies, but does include the relationship between the horizontal temperature contrast and the winds aloft over that zone of contrast. That is more obvious in climatic data than in daily weather.

ORIGIN AND DESCRIPTION

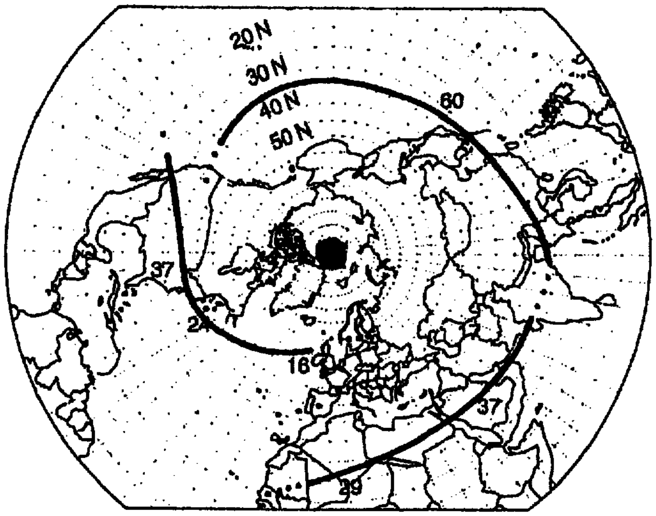
The basic equations of motion require that, once above the friction layer, the change of wind velocity with height must be proportional to the horizontal temperature gradient (contrast). Thus, the higher one goes in this region of contrast, the stronger will be the winds. Indeed, the upper westerlies of both Northern and Southern Hemispheres exist because the poles are colder than the equator. Above about 10 km elevation the south–north temperature gradient tends to reverse, so that the winds decrease upward above that level. Consequently, the level of the jet stream core of maximum winds is generally at 10–13 km above sea level.

Thus, one expects strong winds above the regions of the earth where there is a strong temperature gradient. In the European sector, this means a jet stream in the Mediterranean region where Europe to the north is much colder than Africa to the south, and in the Scandinavian–Baltic area where the air flowing into Europe from the Atlantic is warmer than the cold Arctic air flowing from the ice covered Arctic Ocean. The former position has far more temperature contrast than the latter. Similarly, in the North American sector, the land–Gulf of Mexico and Arctic Ocean–land contrasts give two regions of high probability of strong westerlies aloft. Again, the lower latitude jet stream is the stronger, climatically. The locations of these core areas change with the seasons.

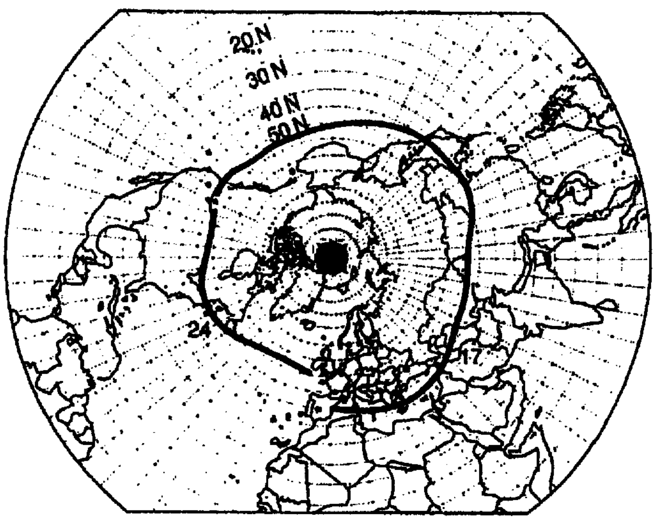
Over the Atlantic and Pacific Oceans there is essentially only one such region of strong temperature gradient each. These lie roughly over the Gulf Stream and Kuroshio oceanic warm currents, respectively. Consequently there is not a continuous or multiple or single jet stream running around the Northern Hemisphere (Figure 1). The numbers along the heavy line give the mean wind speed in meters per second at maxima and significant points.

In January, and the winter season in general, a westerly jet is found over the southern United States and off Southeast Asia, with maxima near the southeast coasts. In crossing the oceans, both shift northward, the Asian jet less than the North American jet, and in the mean they weaken near the eastern edge of the oceans. The main Eurafrican jet starts near the Atlantic over North Africa, weakens in the vicinity of the Himalayas and India, and then strengthens as it becomes the southeastern Asian jet. Thus, the jet stream is not one continuous band of high winds about the Northern Hemisphere.

In July, the jet axis comes closer to being a single river of air but the winds are weaker and more indefinite in position near the west coasts of Eurasia and North America.



Jan 300 mb



Jul 300 mb

Figure 1 The heavy lines show the position of the axis of the jet stream at 300 mb. The values are mean wind speed in meters per second. By far the strongest westerlies along the jet axis in January are off Japan

The Southern Hemisphere has corresponding jet streams, but in that largely water hemisphere the pattern is simpler. There are also segments of weaker, easterly jets at times and in places where the equatorial region is cooler than the tropical region somewhat to the poleward, e.g., subsaharan Africa.

HISTORY

The first observations of large-scale features of the atmosphere with winds in excess of what might occasionally be observed at the ground came with the development of high-flying aircraft. In the mid-1930s there were some encounters with winds on the order of 80 m s^{-1} or less. With the advent of the Second World War, aircraft in Europe occasionally met even stronger winds, but there was little notice taken in the meteorological community at large, possibly due to restrictions in access to information.

During the same time period the Japanese became aware of high winds aloft. With the advent of high altitude bombing missions over Japanese targets in the fall of 1944, a different wind problem arose than in Europe, and produced widespread interest and discussion.

As shown in Figure 1, there are two main locations for branches of the jet stream in Europe: one in the Mediterranean–North African latitudes and one in the Scandinavian area, both with monthly average wind speeds ranging up to 40 m s^{-1} in the south and less than half that in the north. Most aircraft operations were between the two branches. On the other hand the monthly mean speeds in eastern North America and in the Pacific near Japan are 50–100% greater, and the aircraft operations in the Pacific crossed the jet stream position nearly every day in the cooler months. Winds of nearly 120 m s^{-1} were observed. This was sufficiently close to the speed of the aircraft to pose serious military problems. One might say that this was when the jet stream as a large-scale, significant feature of the atmospheric circulation was discovered (Bryson, 1994).

Immediately after World War II a great deal of research effort was devoted to the study of this newly recognized large-scale feature of the atmospheric circulation (Reiter, 1967).

SIGNIFICANCE

The jet stream, lying above the greatest temperature gradients, i.e., the fronts, is therefore the approximate path of frontal cyclones. However, the fast-moving winds of the jet stream must follow the laws of air dynamics and thus do not follow the fronts in detail. The daily adjustment of the one to the other is a large part of mid-latitude meteorology. Climatically speaking, the concordance of mean frontal position and mean jet stream position is much better, as indicated previously. Changes of mid-latitude climate and changes in the mean position of the jet stream are therefore closely related. For example, periods when the hemisphere is warmer are times when the jet stream and associated fronts are generally farther poleward than in cooler times.

See also: **Atmospheric Motions**, Volume 1.

REFERENCES

Bryson, R A (1994) The Discovery of the Jet stream, *Wisconsin Acad. Rev.*, 40(3), 15–17.

Geer, I W, Ginger, K M, Moran, J M, Hopkins, E J, Weinbeck, R S, and Smith, D R, eds (1996) *Glossary of Weather and Climate*, *Am. Meteorol. Soc.*, Boston, 1–272.

Reiter, E R (1967) *Jet Streams*, Doubleday, New York, 1–189.

Figure 1 Photo of a Tornado taken on 5th of June, 1999, southwest of Bassett, NE. (Copyright Howard B Bluestein)

then make impact. Although the pressure inside a tornado can be as much as 10% less than that of its environment, most damage to buildings is not caused by rapid pressure changes. The safest place to be when a tornado strikes is underground. If it is not possible to get to an underground shelter, one should go to an interior closet or bathroom and protect one's head from flying debris.

It is very difficult to determine whether or not the frequency of occurrence of tornadoes or their geographic locations has changed on decadal time scales, even in the US where they are most frequently observed. Reports of tornadoes are heavily weighted by population density, and have been affected by the recent public awareness of weaker tornadoes, which were not reported as often in earlier years. Estimates of tornado intensity using the F scale are not very reliable when the nature of structures damaged by the tornado is not well known; furthermore, many tornadoes strike unpopulated areas where little if any damage can be found.

Estimating future trends in tornado frequency and intensity requires knowledge of why some convective storms produce tornadoes, while others do not. One could estimate the likelihood that the frequency of occurrence of supercells would change from a predicted change in the vertical wind-shear profile and potential buoyancy. The former depends on the nature of the general wind circulation and the latter depends on the vertical distribution of temperature and moisture; both the former and latter might be sensitive to global climate change.

FURTHER READING

Bluestein, H B (1999) *Tornado Alley: Monster Storms of the Great Plains*, Oxford University Press, New York.

Doswell, III, C A and Burgess, D W (1988) Some Issues of United States Tornado Climatology, *Mon. Weather Rev.*, **116**, 495–501.

Fujita, T T (1973) Tornadoes Around the World, *Weatherwise*, **26**, 56–62, 78–83.

Fujita, T T (1987) *US Tornadoes: Part 1, 70-year statistics*, University of Chicago, Chicago, IL.

Kelly, D L, Schaefer, J T, McNulty, R P, Doswell, III, C A, and Abbey, Jr, R F (1978) An Augmented Tornado Climatology, *Mon. Weather Rev.*, **106**, 1172–1183.

Schaefer, J T, Livingston, R L, Ostby, F P, and Leftwich, P W (1993) The Stability of Climatological Tornado Data, in *The Tornado: its Structure, Dynamics, Prediction, and Hazards*, eds C Church, D Burgess, C Doswell, and R Davies-Jones, AGU Geophysical Monograph 79, 459–466.

Trade Winds

The surface winds over most of the tropics are from the east and directed towards the equator. These are the (easterly) *trade winds*. The trade winds are much steadier in both speed and direction than the surface winds at higher latitudes. The mechanism maintaining the trades is understood as a consequence of the existence of the Hadley Circulation (*see* **Hadley Circulation**, Volume 1), which operates as a heat engine to transfer heat away from the warmer low latitudes to the cooler high latitudes. Because of momentum conservation, the rotation of the Earth causes the trade winds to blow from east to west.

See also: **Atmospheric Motions**, Volume 1.

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Transient Response

In climate change research, the term *transient response* is used to indicate the time-dependent character of how a system (typically a numerical model) responds to a perturbation. Because it is sometimes difficult to specify the time-dependent changes in forcings or other boundary conditions, initial studies often are designed to determine the response to a sustained perturbation of some arbitrary size (e.g., doubling the atmospheric carbon dioxide (CO_2) concentration). Such a response indicates the equilibrium response of the system (*see* **Equilibrium Response**, Volume 1). However, because the forcing factors are likely to be changing at the same time other factors are changing (e.g., the CO_2 concentration is slowly increasing at

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Walker Circulation

The atmospheric circulation near the equator varies strongly in longitude over scales of thousands of kilometers. A useful but perhaps oversimplified conceptual model of this variation is one of overturning cells in the longitude/height plane, along with associated patterns of surface temperature, sea level pressure and precipitation. The name given to this conceptual model is the *Walker Circulation* in honor of Sir Gilbert Walker, a Director-General of British observatories in India in the early part of the 20th century, who conducted much research directed at understanding and prediction of the Indian Monsoon. He also identified a number of relationships between seasonal climate variations in Asia and the Pacific region, notably the quasi-periodic variation known as the Southern Oscillation, which is associated with the El Niño.

The current widely used conceptual model and the term Walker Circulation both arise from the work of Jacob Bjerknes in the 1960s (*see* **Bjerknes, Jacob**, Volume 1). The Walker Circulation is comprised of rising motion in regions of high surface temperature (which in turn leads to low sea level pressure and high precipitation coming from deep convective clouds) and sinking motion in regions of low surface temperature (which in turn leads to high sea level pressure and low precipitation). The rising and sinking regions are connected by east/west motions in the upper and lower troposphere, forming the Walker Cells. The east/west motions near the surface are directed from regions of high to low surface pressure, and motions in the upper troposphere at the same longitude are in opposite directions to the surface flows, thereby closing the cells.

The Walker Circulation varies seasonally, but in the mean it has three important rising regions, located in the western Pacific near Indonesia, in equatorial Africa, and over equatorial South America. Following the Pacific cell as an example, the low-level, easterly trade winds form the base of the Walker Circulation, bringing warm moist air from over the Pacific Ocean towards the Indonesian region. Here, moving over normally very warm seas, intense convection takes place, and moist air rises to high levels of the atmosphere. The air then travels eastward before sinking over the eastern Pacific Ocean. The rising air is associated with a region of low air pressure, towering cumulonimbus clouds and rain. High pressure and dry conditions accompany the sinking air. Because this circulation takes place near the equator, the Coriolis force is absent or weak, and a direct zonal (i.e., east–west) circulation spanning the ocean is possible (*see* **Atmospheric Motions**, Volume 1).

Shifts in the location of the Indonesian rising region and the associated large scale changes in the Walker Circulation are a key feature in interannual climate variability associated with El Niño/Southern Oscillation, (*see* **El Niño and La Niña: Causes and Global Consequences**, Volume 1). The strength of the Walker Circulation over the Pacific is conventionally measured by a simple Southern Oscillation Index (SOI), which is essentially the difference in atmospheric pressure between Tahiti and Darwin. During El Niño years, the SOI becomes negative, the Walker Circulation is weakened or even reversed, with suppressed convection in the west and enhanced precipitation in the east over the warm waters characteristic of the El Niño. Positive values of the SOI are associated with stronger Pacific trade winds, warm sea surface temperatures in the western Pacific, and enhanced precipitation in Indonesia and eastern and northern Australia.

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Water Cycle

see Hydrologic Cycle (**Volume 1**)