Glaciers *see* Landform monitoring

Global circulation

The Envelope of Air and Ocean Surrounding the Earth

Thickness Compared with the Dimensions of the Earth

The shape of the earth is very nearly spherical with radius approximately 6400 km. The depth of the oceans averages about 4 km and about two-thirds of the mass of the atmosphere lies within 10 km of sea level. This means that the atmosphere/ocean (the geophysical fluid system) is only a very thin film of material moving and often swirling about the surface. The geophysical fluids are differentially heated by the sun, the tropical regions receiving much more solar heating than the polar latitudes. This differential heating, which takes place mostly through heating the earth's surface, sets up air currents that overturn the fluid and redistribute the heating toward the poles. The earth also rotates about an axis with a fixed orientation in space. This rotation causes apparent forces on the geophysical fluid elements, deflecting them and leading to the formation of circulation patterns, which are the main subject of this entry.

The Interplay of Water in the Climate System

Water substance plays an important role in the climate and circulation system of the atmosphere (*see* **Meteorology**). First, water is abundant at the surface from the oceans, **lakes, rivers, canals and estuaries** and wet ground. As heat is added to the surface much of it goes into evaporating water from liquid to vapor form with negligible change of temperature. The evaporation of water at the surface moderates the temperature at the surface, preventing great temperature swings from day to night and summer to winter. The water evaporated at the surface is recondensed a few kilometers above the surface as the droplets of clouds and precipitation are formed. This process releases heat and warms the middle part of the troposphere where the heat is more easily transported to distant parts of the atmosphere.

The Role of the Circulating Components

In both the atmosphere and the ocean the circulation serves to move heat from the excessively heated tropics toward the poles. As convection lifts warm surface air aloft it begins the complicated process of achieving this goal. In the tropics this is partly accomplished by an overturning cell stretching from the equatorial region to about 30 degrees latitude north and south of the equator. In the middle latitudes westerly winds prevail and embedded in these are the weather systems of swirling eddies of high- and low-pressure areas. It is a remarkable fact that these features, which persist for days at a time, have roughly the same horizontal size passage after passage. In the Northern Hemisphere (NH) the 1000-km-sized high-pressure air masses spin clockwise and the slightly smaller low-pressure areas spin counterclockwise (cyclonic). In the Southern Hemisphere (SH) there are also westerlies, but the highs and lows spin oppositely from those in the NH. High-pressure air masses are typically clear and lows are typically cloudy. These systems spin along their way and bring about the main feature of the general circulation – warm air masses move poleward and cold air masses equatorward. It is a curious fact that these features commonly observed by everyone on television daily are often of about the same size and occur with such regularity yet no passage is ever quite like another.

Absorbed and Emitted Radiation Fluxes

Artificial Satellites

Since the mid-1970s man-made satellites have been measuring the radiation emitted by the earth, the amount coming from the sun and the amount reflected back to space by the planet. We now have reasonable estimates of these components of the radiation budget of the planet (*see* **Remote sensing**). As will be seen below these measurements lead us to the conclusion that substantial quantities of heat must be transported by the fluid media covering the surface of the planet.

Radiation Budget of the Planet

The earth as a material body emits thermal radiation to space roughly at 255 K, a value somewhat lower than the average surface temperature which is about 288 K, since the effective altitude of the emission is several kilometers above the surface where the air is cooler. This leads to an average emission rate of approximately 240 watts per square meter of surface area ($W\,m^{-2}$). The value is some $20\,W\,m^{-2}$ larger at the equator and perhaps $40\,W\,m^{-2}$ less in the cooler polar regions. If there were no external heating source, the planet would cool to a completely frozen surface in a matter of only a few years.

The sun comes to the rescue by sending a steady stream of mainly visible radiation to the planet. The amount of energy reaching an area perpendicular to a ray from the sun is (averaged through the year) about $1370\,W\,m^{-2}$, of which 30% is reflected directly back to space by cloud tops, bright surfaces, etc. The remaining $959\,W\,m^{-2}$ are shared by the whole surface area (four times the area of the intercepting cross-sectional, disk-shaped area), leaving about $240\,W\,m^{-2}$ on the average to heat the planet. Hence, on a global basis the rate of heating just balances the rate of cooling. However, the absorption rate at the equator is roughly $50\,W\,m^{-2}$ more than the global average and at the poles it is some $120\,W\,m^{-2}$ less. The difference in tropical and polar emission rates to space are not enough to balance this strong equator-to-pole gradient. This means that the material in each latitude band is not in thermal equilibrium with respect to radiation. The explanation of this discrepancy lies in the need for heat to be transported along the surface of the earth by the oceans and the atmosphere from the equatorial regions toward the poles.

Atmosphere/Ocean as a Heat Engine

The atmosphere/ocean system of the planet is a heat engine drawing energy from the warmer portions of the planetary surface and converting this heating to kinetic energy, depositing some heat at cooler parts of the planet. The work performed is eventually dissipated in the form of friction. The movements of the fluid media covering the planet form the complex of motions we call the global circulation. The convection process is strongly modified by the fact that the earth is rotating, a complication that causes the circulation to be very different from what might happen on a nonrotating sphere or a very slowly rotating sphere like Venus.

The Atmospheric Circulation

The Coriolis Force and the Deviation of Particle Trajectories

As parcels of air move in the NH they will experience a force perpendicular to their movement and to the right (in the SH it is to the left) called the Coriolis force. This 'force' is actually not a real force but rather an artifact arising because the planetary surface rotates and Newton's laws of mechanics apply to inertial reference frames (which are not rotating). The Coriolis force is simply a correction to take this into account. Of course, an air parcel or any other moving body will experience this force and be accelerated when observed from a reference frame fixed at the surface of the rotating planet. The Coriolis force is largest at poles and is zero at the equator.

Stratification of the Atmosphere/Ocean Media

Another important feature of the atmospheric and oceanic systems is that they are stratified such that more dense matter is overlain by less dense matter. In the case of the atmosphere, the density is roughly exponentially decreasing with an exponential scale height of about 10 km. The vertical gradient is less in the ocean, but roughly speaking each medium consists of a system of stacked layers with the denser members below. Such a configuration is said to be stably stratified.

Even a medium stratified in a stable manner when at rest will become unstable if the heating below is enough to make the lower layers sufficiently buoyant, or if the horizontal velocity above is sufficiently greater than that below to cause a shearing instability. This kind of instability is common in the middle latitudes in the atmosphere and is partially responsible for the stormy weather at those latitudes. It is a remarkable feature that even though the atmosphere has a characteristic depth of about 10 km, the horizontal sizes of eddies (highs and lows) in the middle latitudes are about 1000 km or about one-sixth the radius of the earth.

Mean Quantities and Eddies

Zonal Averages: Mean Circulation. If we average the mean circulation pattern over a long time and over longitude we arrive at an important measure of the atmospheric circulation: the zonal mean circulation. Figure 1 presents a slightly idealized cross-section of the zonal wind component (west to east) under mean annual conditions. The most prominent feature is the strong jet at about 35 degrees latitude of each hemisphere – these are known as the subtropical jets. The units on the ordinate of the figure are mbar or millibars (one-thousandth of standard atmospheric pressure). These jets flow from west to east in each hemisphere and are very intense at their cores (about 25 m s^{-1}). Note that the winds flow from east to west throughout most of the tropics.

Figure 2 shows a similar cross-section for mean annual and zonal averaging, but of the mass stream

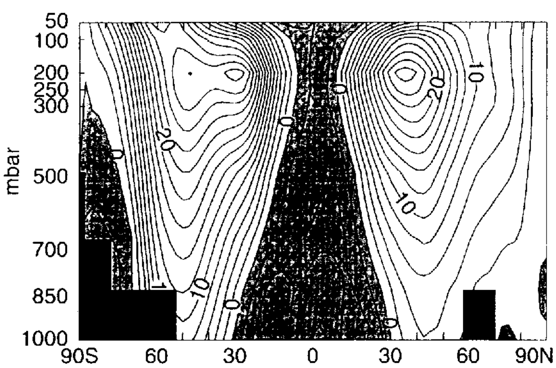

Figure 1 Contours of the zonal mean of zonal wind speed (unit: m s^{-1}). The ordinate signifies altitude described by atmospheric pressure (unit: mbar). Positive values indicate westerly winds and negative values (shaded regions) easterly winds. Adapted from Figure 2.6 of H.B. Singh (1995) with permission from Van Nostrand Reinhold

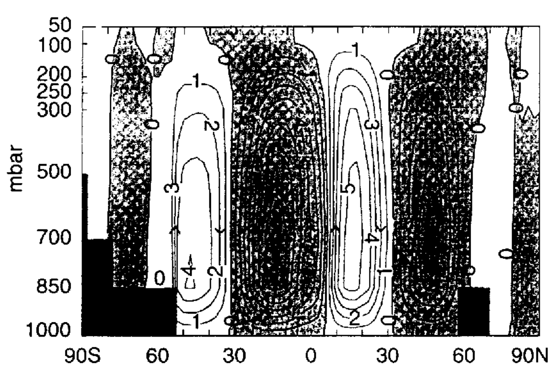

Figure 2 A cross-section of the mass stream function for the meridional circulation (Unit: 10^{10} kg s^{-1}). Adapted from Figure 2.10 of H.B. Singh (1995), with permission from Van Nostrand Reinhold

function. The contours in this diagram indicate the average speed and local direction of the circulation cells. The important feature here is the circulation cell in each hemisphere near the equator. In each case there is strongly rising air near the equator approaching the very top of the atmosphere before turning polewards in each hemisphere – these cells are known as the Hadley cells after the English scientist who first suggested their existence in 1735. The rising branch of the Hadley cells is near the equator and forms a band encircling the planet known as the intertropical convergence zone (ITCZ) – it is very prominent in images taken from geosynchronous satellites. As the seasons progress through their cycle, the ITCZ moves back and forth across the equator, with the Hadley cell on the winter side becoming stronger by roughly a factor of five over its counterpart on the summer side. The winter side cell is also about 50% larger in latitudinal extent.

Other features to be noted in the two figures are the weaker cells in the middle latitudes known as the Ferrel cells after William Ferrel who published his theory in 1856. They are much less prominent than the Hadley cells and therefore require more data to verify. As will be indicated below, the situation in the middle latitudes is much more tangled with unstable motions and wave-like structures stretching around latitude circles, so that the Ferrel cells do not help much in understanding the motion in this band.

Departures from the Means. The zonal mean motion gives only a partial picture, since much of the interesting activity of the flows is averaged out in constructing such a diagram. For instance, the steady wave-like structures around latitude belts, known as stationary eddies, are excited by features in the topography and land–sea distributions.

In addition to stationary eddies there are constant fluctuations of the winds about the means which are time-dependent – these are called transient eddies. The transient eddies are mainly generated by the instabilities in the atmospheric flows. The largest fluctuations appear near the jet streams where the shearing of flowing air becomes unstable, often leading to meandering and splitting motions of the jets.

Both stationary and transient eddies of wind are interesting because the zonally and temporally averaged total kinetic energy per unit mass (half the wind speed squared) can be decomposed into a sum of contributions from the mean motions, stationary eddies, and transient eddies (similar to an **analysis of variance** decomposition). These individual contributions can be measured from observed winds and can be calculated from computer models of the general circulation for comparison. We are also interested in the apportionment of various transports by the eddies vs. the mean motions.

Conserved Quantities and the Equations of Motion

Mass. Since matter is neither created nor destroyed as it moves about in the atmosphere, the mass of fluid substance is conserved along the motion of packets. This idea can be expressed quantitatively as a partial differential equation called the equation of continuity. It is fundamental to any problem involving fluid motion. The simplest expression of it is in pipe flow where, as mass is pushed through a pipe at a certain rate, some constraints on the speed of the local flow are implied. In this simple case if the fluid is incompressible, a constriction of the pipe will lead to larger speeds as the fluid passes through that portion of the pipe.

Water Substance. Water vapor and water droplets are carried along by the flow of air except for the eventual settling out of the larger particles by gravity (drizzle droplets are about 1 mm in size). The sum of water vapor and droplet mass is conserved in an air parcel as it moves about in the atmosphere. The vapor may convert to liquid or solid form along the path but no mass is lost except of course by precipitation. There is an equation of continuity for water substance that can be added to the mass conservation law of the last paragraph. This is usually written in the form of the water vapor concentration or the mixing ratio (the number of grams of water substance per kilogram of air). As an air parcel moves along its path the mixing ratio of water vapor is conserved. If the dew point is reached by some kind of cooling, water droplets will form on the always available condensation nuclei.

Conservation of Other Species. The conservation of other chemical species can also be accounted for by similar continuity or budget equations. Basically, these equations state that the rate of gain in concentration of a species along its path is the difference in the rate of production (sources) and the rate of loss (sinks). In the case where chemical reactions can take place there will be several equations coupled together with gains of some species corresponding to losses of others as in elementary chemical kinetics (*see* **Atmospheric dispersion: Chemistry**).

In the ocean a primary determinant of density is the salinity of the water. Salinity can change, especially near the surface of the ocean where evaporation (enhancing salinity) or precipitation (diluting salinity) can take place. An interesting phenomenon occurs near the margin of sea ice where the formation of new ice leads to a rejection of salty water. This briny residue can significantly increase the density of the water and in conjunction with the greater density of cold water can cause sinking to very great depth.

Momentum. According to Newton's second law a mass subjected to a force will be accelerated by an amount inversely proportional to its mass. A fluid such as the ocean or atmosphere can be considered to be made up of small elements each of which is separately subject to Newton's second law. The forces that can act on a fluid parcel are as follows. The gravitational force acts to force each parcel toward the center of the earth. The gravitational force in a fluid of variable density gives rise to buoyant forces that tend to stratify the fluid with more dense elements lying below the lighter ones. The Coriolis force acts at right angles to the velocity of the particle (to its right in the NH and to its left in the SH). At horizontal speeds typical of atmospheric and oceanic parcels this is an important force causing significant deviation of parcel trajectories over a period of days.

Gradients of pressure cause a force on parcels. The force is from regions of high pressure toward those of low. For example, there is a large gradient of pressure downwards in the atmosphere, leading to a vertical force. This force is nearly exactly balanced by gravitational force on a parcel. There are also frictional forces in the atmosphere and ocean. These become important near boundaries such as in the flow of the atmosphere over a rough surface (*see* **Atmospheric dispersion: Complex terrain**). The friction prevents slippage of the flow at the interface imposed by the boundary, but this very thin molecular boundary layer of only a few millimeters causes turbulence and the momentum transfer is transmitted to larger scales (eddies) through turbulent motions that are intense near the surface, diminishing in intensity with distance (*see* **Turbulent diffusion**). Such turbulent motions form a layer of turbulent air known as the planetary boundary layer, typically 1–2 km thick. The ocean also has a turbulent boundary layer near its surface due to the interaction of the winds above. It is called the mixed layer of the ocean and is about 50–100 m thick depending on location and season. Boundary layers also form in the horizontal directions near shorelines or continental margins in the oceans (see the discussion below of the Atlantic Gulf Stream or the entry on **coastal processes**). The last force that can in principle be included in the expression of Newton's second law is viscosity, the internal friction. Viscosity is important at very small scales, typically 1 mm in the atmosphere. We must recognize that the term in the equation that represents viscosity itself is negligible in atmospheric and oceanic flows, but its presence is the ultimate cause of the formation of boundary layers and other turbulent features.

Energy and Thermodynamics. The sum of energy in all its forms for all the parcels of air is conserved in atmospheric and oceanic flows. These forms transform energy from one type to another and from one parcel to its neighbors along the motion. The forms of energy for a parcel include its kinetic energy, its gravitational potential energy, and its thermodynamic internal energy. There can also be a component of internal energy associated with the latent heat of the parcel if it contains water substance. Transfers of one of these forms to another occur often along the trajectory of such a parcel.

The parcel can have significant interactions with its environment that lead to the energy transformations. For example, gravitational potential energy can be transformed into kinetic energy, or the parcel can do work on the surrounding air by expanding. The latter case is a very common occurrence. Consider a parcel that is forced to rise by flow over a mountain. As the flow proceeds up the mountain slope the parcel's pressure changes quickly in such a way as to match its internal pressure to that of its local elevated environment. The lowering of pressure causes the parcel to expand its volume, performing work on the surroundings. The expansion work comes at the expense of the internal energy of the parcel and this leads to a lowering of its temperature. If the temperature is lowered sufficiently to reach the dew point of the moist air, a cloud base will form as the parcel rises. Next the production of water droplets leads to a release of latent heat which can cause the parcel to become more buoyant than the surroundings. The added buoyancy brings on an additional gravitational impulse in the vertical direction.

A parcel can be heated by several mechanisms. These mechanisms include direct contact with an adjacent body (such as the ground) through heat conduction, warming by the absorption of visible solar radiation, warming or cooling by the exchange of infrared radiation with other material bodies within the earth/atmosphere/ocean system, or the parcel can be warmed by the latent heat released to sensible heat from the condensation of vapor to liquid or solid form – the reverse can also occur. Chemical reactions can also occur in the parcel to warm it. These forms of heating or cooling a parcel are referred to as diabatic heating. Many atmospheric motions occur on such a short time-scale that the radiative heating and cooling can be considered negligible compared to the exchanges due to work done by the parcel on its environment, or vice versa.

The expression of the conservation of energy in the equations governing the motion of the atmosphere and ocean is the first law of thermodynamics as expressed for an individual parcel along its motion. Equated to the rate of change of thermodynamic enthalpy (the heat capacity at constant pressure multiplied by the rate of change of temperature along the path) are the various sources and sinks of enthalpic energy, including diabatic heating rates and the rate of pressure work done by the parcel (volume multiplied by rate of pressure change along the path).

In transforming energy from potential to kinetic one must take into account that there is always a certain amount of potential energy in the atmosphere simply because of its vertical extent. Disturbances of an atmospheric column's thermal profile from its equilibrium vertical stacking lead to the possibility of potential energy which is available to create kinetic energy. Since not all the potential energy is available to produce kinetic energy it is necessary to distinguish this concept of available potential energy.

Angular Momentum. The angular momentum of an isolated body under no net torque will be constant. The angular momentum of the earth is naturally much larger than that of the ocean/atmosphere system. If the solid earth and the geophysical fluids are considered as separate masses, they can exert a net torque on one another. But since there is no evidence that there is a secular trend (a steady trend independent of seasons or other internal periodic changes) in the angular momentum of the geophysical fluid system, it is a good assumption that the net torque exerted by the atmosphere on the solid earth including the ocean is negligible.

In considering the general circulation of the atmosphere this constraint of no net torque is an important feature to include in any attempt to construct a consistent picture of the flows. Locally there are many such torques exerted. For example, even over ocean surfaces the trade winds tend to blow from east to west in the tropics. This frictional contact will exert a torque tending to slow the rotation of the solid earth. This torque is compensated by the westerlies of middle latitudes. It is interesting that there can be seasonal deviations of the torque exerted by these winds since the seasons can make for swifter westerlies in the winter hemisphere. Also the conditions present during El Niño, which will be explained later in the entry, can cause changes in these and other circulation related torques. Both the seasonal cycle and the fluctuations during El Niño can be detected by very accurate (astronomical) estimates of the length of day. These changes are of the order of milliseconds but are measurable with modern satellite techniques.

Circulation Variables of Interest. The main variables of interest in treating the circulation are the wind components: westerly, southerly, and vertical. The vertical component of wind is usually two to three orders of magnitude smaller than the horizontal components, especially in the middle latitudes and away from storm centers. The fields of temperature and moisture are also important in determining the circulation. It is not only possible but also convenient to decompose each of these variables into a mean and a fluctuating part. The fluctuating part might be a departure from the zonal mean or a departure from a temporal mean or both. In so doing it is possible to understand how heat or other quantities are transported across latitude circles. That is to say, we can distinguish between heat being carried by mean motions (for example the large-scale overturning in the tropics), as opposed to the transport by eddies (for example the passage of heat polewards by highs, lows, and the associated frontal systems in the middle latitudes). Other important distinctions can also be made, such as the decomposition of the kinetic energy of the atmosphere into its mean and eddy contributions. These decompositions lend insight into the nature of the flows and the way certain quantities are moved from one place to another by the geophysical fluids. For example, one might wish to know whether most water vapor is transported by mean motions or by eddies. It turns out that in the tropics water vapor is transported primarily by the mean flow near the surface, but as the middle latitudes are approached, eddy motions begin to carry most of the water vapor.

Some Approximately Conserved Quantities. Along the path of a parcel of air there are some quantities with values that are preserved. In some cases the conservation is only approximate, but sufficiently accurate that it is useful in qualitative descriptions. Sometimes the approximation is good enough for accurate theoretical and numerical descriptions. The first is associated with the movement of an air parcel in simple vertical excursions. We assume the parcel's internal pressure quickly equalizes with that of the local environment. But such a parcel, whether in motion by its own buoyancy or by being forced by the flow (say, up a mountain slope), does not absorb heat from its surroundings. Such flow is called adiabatic. Considering the parcel to be a thermodynamic system, its exchanges with the environment can also be considered reversible and therefore an adiabatic flow is also isentropic, i.e. no change in entropy. Even large-scale horizontal flows are approximately adiabatic and this is a useful approximation in theoretical studies.

Vorticity is a measure of the spin of an air mass. It is roughly the line integral of the tangential velocity component of the flow along the perimeter of the geographical region enclosing the mass (called the circulation) divided by the area of the region. In a fluid of constant depth the vorticity of an air mass will be conserved along its motion unless torques in the form of friction at the surface are applied (angular momentum conservation). For air masses on a rotating planet the local vorticity associated with the planetary rotation must be added to the vorticity relative to the planetary surface. This absolute vorticity is conserved if the depth is fixed.

If the depth of the fluid is not fixed there is still a principle called the conservation of Ertel's potential vorticity. This conservation principle takes into account the depth locally and affects the local moment of inertia (recall the analogy of a rotating ice skater who pulls in her arms to increase her rotation rate). It is possible to combine the concept of adiabatic flow with the conservation of angular momentum for a parcel. This is the concept of conservation of potential vorticity. Many approaches to understanding the general circulation make use of the conservation of potential vorticity.

Tropical Circulation

Hadley Circulation. The most important components of tropical circulation are the Hadley cells. Zonal averages of the vertical and meridional winds reveal a net overturning of the atmosphere, rising near the equator and sinking at about 30 degrees latitude in both hemispheres. The Hadley cell is an example of a direct circulation because it is the direct response to differential heating from the tropics to the subtropics. Most of the heat at the surface in the tropics is transferred to the air flowing overhead by evaporation of water from the surface. The water vapor is then carried to great heights in the equatorial belt by tall (20–50 km) towering cumulus clouds, converting to droplets and finally precipitation. At these high altitudes the flow is carried away from the equator toward the middle latitudes to sink in the neighborhood of 30 degrees latitude.

It is interesting that the return flow along the surface is very shallow, confined to the lowest kilometer or two in the so-called planetary boundary layer. This return flow is called the trade wind. It is steady and rather intense, of the order of $10\,\mathrm{m\,s^{-1}}$. As the trade winds flow over mostly tropical oceans they become laden with moisture from the warm waters below. Small cumulus clouds pepper the trade wind zones. As the trades converge near the equator they rise into an irregular row of tall towering clouds delivering the heat evaporated from the surface many kilometers distant directly to the convecting air.

The trade winds do not blow strictly parallel to the meridians. In the NH they blow from northeast to southwest. This can be explained by an argument based upon the conservation of angular momentum. Consider a ring (torus) of air at the surface at 30 degrees north. The radius of the ring is to be enlarged as it approaches the equator since the ring's radius is about 13% larger at the equator. The moment arm is therefore larger and thus the moment of inertia is larger. This process will cause the angular velocity of the air in the ring to decrease. The equatorward moving ring will appear to have speeded up relative to the earth's surface, creating the northeast to southwest flow. Actually the wind speed is considerably slowed by friction near the surface, but the concept is valid nonetheless. The same argument can be reversed for air aloft which is flowing northward. It will be speeded up by the contraction of its radius and will therefore be seen to flow from southwest to northeast. A similar result can be arrived at from consideration of the deflection due to the Coriolis force.

The near surface return flow from both the northern and southern branches of the Hadley circulation converge on the equatorial belt from east to west. This has an important consequence for the wind stresses on the ocean surface, pushing water along the equatorial belt from east to west. Since the tropical oceans are confined into basins with western boundaries, this leads to a buildup of water mass at the western end of such basins. The consequences of this buildup lead to the so-called El Niño oscillation of the climate system.

While the schematic view of the zonal mean Hadley circulation is very useful from a theoretical point of view, it must be kept in mind that there are large deviations from the zonal mean from place to place and from time to time. For the most part the circulation occurs locally in bursts and the equatorial belt of rising air is not a simple circle but patchy, wavy, and irregular. It is much more intense over land than over ocean. The Hadley circulation has a large seasonal cycle which will be discussed below.

The ITCZ. The ITCZ is the belt around the tropics that makes up the rising branch of the Hadley cells. Most tropical precipitation falls in this region. In fact, the descending branches of the Hadley circulation contain almost no precipitation and are homes of the great deserts of the world, such as the Sahara and the interior of Australia. Those parts of the ocean that lie below these descending branches are likewise without precipitation – the Bermuda High, for example. By way of contrast much of the world's rain forest lies beneath the ITCZ – for example, the Amazon Valley, the Congo River Valley, and the East Indies. The ITCZ has been called the 'firebox' of the engine driving the general circulation of the global atmosphere. The precipitation intensity varies greatly as one traverses around the equatorial belt.

It is greatest over land masses because the heating of land by the tropical sun raises temperatures more than over oceans, increasing evaporation and sensible heating. But there are fluctuations even in relatively homogeneous regions, such as over the broad Pacific Equatorial Basin. For example, waves of precipitation (the so-called Madden–Julian waves) move from west to east such that it takes 30 days or so for such a wave to encircle the planet.

The Monsoon Circulations. The midday sun moves across the latitude circles as the seasons march. At summer solstice it is well into the NH at 22.5 degrees latitude. At the equinoxes it is at the equator and so on. Without the atmosphere and oceans the warmest spot would be at these latitudes at that particular season. The proximity of ocean will delay the warmest time past solstice by as much as a few months while in continental interiors the delay is only a few weeks. This differential of surface warming is enough to provide a qualitative understanding of the movement of the ITCZ over Africa. Just after northern summer solstice the ITCZ moves to the Sahel area just south of the Sahara Desert, bringing the rains and the greening of the arid regions, followed by the migration of great herds of wild animals. As summer progresses the ITCZ swings southward again crossing the equator just after equinox. At the equator there are two rainy seasons per year, one as the sun passes going south and another as it goes by on its way to the north. At the extremes of the seasonal excursion there is only a single rainy season, at the peak of summer. When the rains come they are mostly from tall cumulus clouds with torrents of downpour. The swing of the ITCZ with all its accompanying trade wind patterns on its seasonal cycle is called the monsoon.

India depends on the monsoon every year for its precipitation. As the subcontinent heats up in spring to very high temperatures the ITCZ swings north of the equator and behind it the surface winds come across the equator bending to flow from southwest to northeast. The monsoon rains begin to fall toward the end of May in the southern part of India and move progressively into the interior as the summer season proceeds. The sky is cloudy most of the time between rains and the air cools to a bearable level again. Most days have heavy rain. The flow of the winds and the pattern of the deep convection are strongly influenced by the land–sea geography around India, and even the high elevations in the north are important in the details of the circulation. Finally in late summer the rains move again to the south and India begins its long dry season. The monsoon is variable from year to year and sometimes the floods are huge with loss of life and property (*see* **Natural disasters**). At other times the monsoon fails to produce the expected amounts of rain and shortages of food occur.

The monsoon is a feature of the entire southeast Asian subcontinent. It sweeps back and forth across the East Indies twice per year and into Indochina and northern Australia on its extremes. The effects of the monsoon are felt as far north as Taiwan and Korea through indirect circulation and frontal movements. The monsoon circulation of Asia is influenced by global phenomena such as El Niño, which will be discussed below.

Walker Circulation. A rather peculiar tropical circulation pattern was discovered by Gilbert Walker during the 1930s. There is a rising of the air over the East Indies and a subsequent sinking along the equator in the Mid-Pacific; it has been called the Walker circulation. Walker discovered a modulation of this circulation, a seesaw pattern of pressure between Darwin, Australia, and Tahiti, which is located in the Central Equatorial Pacific. When the atmospheric pressure at the surface is high in Darwin, it is low in Tahiti, and vice versa. This seesaw oscillation varies from year to year and seems to have extremes spaced two to seven years apart. When convection and precipitation occur abundantly over the East Indies (as indicated by low pressure in Darwin) the air is sinking in the Middle Equatorial Pacific near the dateline (as indicated by relatively high pressure in Tahiti). Low pressure in Darwin means heavy rain and a strong monsoon circulation in the East Indies. In the opposite phase of the cycle, the air is sinking over the East Indies and lifting out over the Middle Equatorial Pacific. This situation leads to drought in northern Australia and, depending on the intensity of the event, potential disaster in countries such as Borneo. The modulation of the convergence and divergence of surface flows is known as the southern oscillation. It is now known to be intimately connected with the phenomenon known as the El Niño cycle. The two together are often referred to as the El Niño–southern oscillation (ENSO).

Easterly Waves. Outside the heavy convergence zones of surface flows the tropics are generally very calm with steady trade winds and occasional afternoon showers especially on tropical islands. Small depressions in the atmospheric pressure are usually quickly eliminated by flows of mass that equalize the pressure. This is more easily accomplished in the tropics than elsewhere because the Coriolis force is much weaker in the tropics, being proportional to the sine of latitude.

The tranquillity of the tropical weather is occasionally interrupted by a weak traveling disturbance moving from east to west (in either hemisphere), called an easterly wave. These disturbances are generally mild, traveling at speeds of 20–40 $\mathrm{km\,h^{-1}}$, and produce some precipitation, but the associated winds are not usually destructive.

Tropical Storms. Tropical disturbances cover a wide range of intensities. These are familiar to most readers because of their thorough coverage in modern news reports. The storm progresses through the stages of development from tropical disturbance to tropical depression to tropical storm. The final stage is the hurricane (so-called in the North Atlantic and the Eastern Pacific – in the Western Pacific they are called typhoons and in the Indian Ocean they are called cyclones; for simplicity we refer to all of these as hurricanes in the rest of this discussion). These strongest category storms graduate to hurricane status when their near surface winds exceed 118 $\mathrm{km\,h^{-1}}$.

In the NH, the hurricane is a large counterclockwise (clockwise in the SH) swirl of winds, with a central disk-shaped region called the eye, which can be 15 km to several hundred kilometers in diameter. The eye, which itself is usually clear and warm, is surrounded by a tall circular wall of heavy rain and high winds which encircle the eye. The hurricane wall can reach 12 km or higher in altitude. Hurricanes always form over warm ocean waters and they draw their energy from this warm water in the form of latent heat release. Analyses have shown that inevitably the air in the wall portion of the hurricane is always such that parcels lifted from the surface and continuing to condense water vapor into cloud droplets will remain warmer than surrounding environmental air, leading to additional buoyancy and lifting.

Hurricanes never form closer than about 5 degrees latitude to the equator, this presumably because of the necessity for sufficient Coriolis force to sustain the cyclonic motion. Hurricanes soon dissipate over land or colder waters. They generally move from east to west embedded in the trade winds with a slight poleward drift, but the proximity of land can distort the paths into sometimes intricately looping trajectories. The formation of hurricanes requires at most weak nonshearing winds aloft but diverging air at high altitudes is necessary. These conditions are to be contrasted with the strong cyclonic storms in the middle latitudes.

North Atlantic hurricanes seem to be inversely correlated with the occurrence of El Niño. This correlation surely has to do with the conditions aloft near the tropopause conditioned by the altered Walker circulation during El Niño episodes.

Middle Latitude Circulation

Geostrophic Flow. As mentioned earlier, the Coriolis force increases with latitude and begins to become very important at latitudes polewards of 5 degrees latitude. Air well above the boundary layer (a few kilometers) experiences negligible friction from the ground and the only horizontal forces acting on an air parcel are the pressure gradient and the Coriolis force. If these are the only two forces acting then a steady flow can emerge with the parcel trajectories moving along isobars with the higher pressure to the right (NH; SH to the left). As a parcel moves along the isobaric path the Coriolis force directed at right angles to the right of the path is exactly balanced by the pressure gradient force directed at right angles to the left of the path. Isobars are seldom along straight lines and instead are often closed curves around high- or low-pressure areas. This causes flows around highs to be clockwise (NH; SH counterclockwise) and the flows around lows vice versa. Flow of this type which balances the Coriolis and the pressure gradient forces is called geostrophic flow. Geostrophic flow is a good approximation to describe most movements of air above the boundary layer in the middle latitudes.

Modern satellite observation on energy radiation reveals a problem in attempting to explain the general circulation: the geostrophic flows cannot transport the heat necessary to explain the heat transfer polewards required by the radiation energy budget (*see* **Radiation and radiative transfer**). Hence, other more subtle mechanisms must be involved in removing heat from the tropical areas and moving it toward the poles. As we will see in the next section, this is accomplished by eddies which form the highs and lows embedded in the westerlies of the middle latitude flows.

Instabilities and Storms. As flows develop there will be differentials of speed across the flow – some elements will be moving faster than those next to them, a phenomenon known as shear (the rate of change of velocity with respect to distance perpendicular to the velocity). When shear becomes large in a flowing fluid the flow becomes unstable. An example is the flow past a flag. At the surface of the flag the air must be nearly stagnant because of friction, but only millimeters away it is moving with the speed of the ambient wind. This leads to unstable flow and the flag begins to flap.

An example more pertinent to the atmosphere is the flow of a density-stratified fluid (larger densities below). If there is a shear such that the fluid above is moving faster than that below the flow will similarly become unstable (turbulent) when the shear exceeds a certain critical value, depending on the stratification. Once the flow becomes unstable, eddies are formed and the motion becomes disorganized and chaotic, i.e. turbulent. Earth offers a further complication, namely its rotation and associated Coriolis force. When thermal contrasts from one latitude belt to another (north to south) increase due to differential solar heating, the geostrophic winds will tend to increase with vertical height, inducing a strong vertical shear. As in the example above there will be a critical value of the shear, related to the horizontal temperature gradient, above which the flow becomes unstable. This instability is called baroclinic instability. Eddies are formed and they move with the westerly flow. In the entanglements of parcel trajectories the strict geostrophic motions are disrupted sufficiently to allow the necessary heat transports to take place.

In the middle latitudes most of the heat transport occurs in the few kilometers just above the boundary layer and it is accomplished by the eddies in the form of high- and low-pressure areas and their associated frontal boundaries. Approximately half the heat is transferred in the form of latent heat and half is in the form of sensible heat. Water vapor is similarly transferred mostly by these same eddies away from the tropics, eventually falling in the form of precipitation.

Seasonality

The Seasons in the Middle Latitudes. As the NH winter proceeds the band of westerlies moves equatorwards by at least 1000 km. Areas that are tropical (no eddies – highs, lows, and associated fronts) in summer begin to experience eddy activity and the first fronts begin to cross. Southern California begins to experience its short rainy season as occasional disturbances dip far enough south to bring weather.

As the season progresses the polar region becomes perpetually dark and the ITCZ (in most places) slips across the equator to the summer hemisphere. This causes the contrast between the 'thermal equator' and the winter pole to become very large. The local gradients of temperature increase and baroclinic instability becomes more frequent and more intense. Winter cyclones begin to dominate the weather in the middle latitudes. Land areas with their smaller effective heat capacity cool faster than maritime areas, sometimes allowing the thermal gradients to become larger over the continents, increasing the incidence of baroclinic instability; this effect can be at its largest near the land–sea boundary of middle latitude continents. Aloft where the instability begins this might be downstream from where the contrasts are at the surface. In fact, many storm tracks intensify in the westerlies just downstream from continental margins and over the sea – this partly because of the contrasting warm sea surface (lagging behind the cooling by several months) and warmer land areas to the west which have caught up with the cooling much sooner.

Hadley Seasonality. The Hadley circulation described earlier undergoes a very strong seasonal cycle. As the ITCZ moves across the equator into the summer hemisphere, the Hadley cell on the summer side weakens considerably. Along with the shifts equatorward of the westerlies in winter is the shift in the poleward edge of the Hadley cell on its winter side. The circulation of mass in the winter cell is many times larger than that in the summer side cell. Much of the polewards heat transport is delivered to the middle latitude westerly eddies by enlarged winter-side Hadley cell. The heat transport in the tropics is accomplished mainly by mean circulation rather than by eddies.

General Circulation Models

The laws of physics as they apply to the atmosphere were described in an earlier section. It is possible to express these laws in the form of partial differential equations, whose dependent field variables depend on position on the globe, altitude, and time. One can then proceed to solve these equations by means of numerical algorithms which can be implemented on a digital computer. The use of general circulation models (GCMs) of the atmosphere and ocean has been a great benefit to advancing our understanding of the atmosphere/ocean circulation problem. The models provide simulated circulation patterns that are quite realistic, although detailed quantitative flaws remain. Nonetheless, the model simulations allow us to experiment with including or excluding certain features to see the consequences – almost transforming the problem into a laboratory science.

The Equations of Motion. As elaborated on earlier, Newton's laws, the equations of state of the gases and liquid mixtures (e.g. salinity in the oceans), and the thermodynamic and material conservation equations all together determine the flow of the atmosphere and all of its transformations. The forces include the pressure gradient force, the gravitational force, and the Coriolis force. The unknowns include the three vector components of wind, the density, the pressure, the temperature, and the water vapor mixing ratio. In addition, one must keep track of the condensation of water vapor and the associated cloud amounts and their vertical and horizontal distributions, and the radiation field with its associated diabatic heating and cooling. The condition of the surface must be kept track of with such variables as surface roughness, soil moisture, vegetation, and snow cover.

As in any mechanics problem the problem is an initial value problem, with initial conditions and boundary conditions to be taken into consideration in starting the numerical solution procedure. In a weather forecast the initial conditions are crucial. One gathers data on all the variables worldwide and after proper smoothing and other technical adjustments are made, these are inserted into the model and a forecast is generated. Weather forecasts based solely on the atmosphere (with the sea surface conditions held fixed) are known to deteriorate to virtually no information beyond **climatology** within a few weeks. This means that we can use any set of initial conditions and wait long enough for the general circulation to establish itself in the simulation. By generating a set of runs of this type in parallel one can look at averages across the runs to learn how the model performs and in turn learn more about the general circulation.

Numerical Considerations. Before proceeding to numerical experimentation with model simulations, many preliminary considerations must be taken into account. A typical global model used in numerical simulations is implemented on a grid spread over the earth. The grid might be roughly rectangular on the surface with box sizes on the order of $100 \times 100\,\mathrm{km}$. Since the earth has a surface area of about $5 \times 10^8\,\mathrm{km}^2$, we will have roughly 5×10^4 boxes to contend with. In addition, we must have about 20 layers in the vertical, bringing the total to 10^6 boxes. This means we have a tremendous bookkeeping job just to evaluate all of the variables on the grid. It takes the world's fastest computers to conduct these kinds of experiments. Finally, numerical considerations suggest that in order for the simulation to proceed without numerical instabilities, the time step must be of the order of a few minutes or less. Hence, to simulate months (or thousands of years in a climate change simulation), much computer time and storage capacity will need to be called upon.

Ocean Circulation

Some Fundamental Ocean Properties

The oceans cover about 71% of the global surface area – the rest is land. Sea water carries a significant amount of dissolved salt, typically 35 parts per thousand, and the ocean's average temperature is near 4 °C (*see* **Oceanography**). The average depth of the world oceans is around 4 km, but there are many ridges and trenches, some dipping to 12 km. Around the perimeters of the continents are the continental shelves that extend an average of about 65 km into the oceans with an average slope of about 1 in 500 – this to be contrasted with 1 in 20 for the slope beyond the shelf.

The density of sea water is controlled by its local temperature and salinity. Open ocean values near the surface are around $1020\,\mathrm{kg\,m^{-3}}$ and at a depth of 10 km it is roughly 1070 kg m^{-3}, compared to the density of pure water at standard conditions of 1000 kg m^{-3}. There is a considerable latitudinal variation in surface salinity, with minima where precipitation is prevalent (e.g. in the ITCZ) and maxima in the subtropical highs (the descending branches of the Hadley circulation) where evaporation exceeds precipitation. The water can be very saline near the margins of sea ice as winter comes on because of the rejection of salt in the freezing process.

The major ocean areas are usually taken to be the Southern Ocean, the Atlantic Ocean, the Pacific Ocean, the Indian Ocean, and the Arctic Sea. Except for the first these are essentially surrounded by landmasses with connecting passageways. There are also a number of smaller bodies of water which are generally connected with the seas, but have characteristics of their own.

The vertical structure is generally stably stratified – i.e. the more dense layers are below less dense layers. The upper 100 m or so of water are generally turbulent and well mixed due to the wind stresses applied above. Very slowly rising cool waters (a few meters per year) throughout the world ocean counter the tendency of turbulent heat conduction from the mixed layer downwards. The result is a layer about 500 m in depth (including the mixed layer) where warmer waters lie above cooler waters below. This depth is called the thermocline. Actually the temperature variation tends to fall exponentially from the surface value to that characteristic of deep water with a characteristic length scale of the thermocline depth. The water below the thermocline tends to be more uniform and the stability is roughly neutral.

Types of Ocean Circulation

The ocean's circulation can be roughly divided into two types, wind driven and density driven.

Wind-driven Circulation. When wind blows over an ocean surface, the surface waters respond by accelerating initially in the direction of the wind. As the current speeds up, the Coriolis force, acting at right angles to the flow velocity, will act to deflect the motion to the right (NH; left in the SH). In addition, as the flow accelerates there will be friction forces that act opposite to the direction of the velocity. The resulting motion will be such that the vector sum of the three force vectors vanishes. The upshot is a flow that is not along the wind but between 0 and 45 degrees to the right (NH; left in the SH) of the wind stress. This is known as Ekman flow, after the scientist who first explained it in 1905. For example, in the NH trade winds, the winds will come from the northeast to the southwest. The Ekman flow will tend to be nearly due west. Hence, in the great basins of the NH, there is a tendency for the flows north of the equator to be from east to west across the basin. Wind-driven circulations tend to be in the horizontal plane tangent to an ocean basin.

Density-driven Circulation. Another easily identifiable circulation type is driven by density differences. In this case more dense water sinks to lower depths in the ocean. In spite of this simple mechanism, the opportunities for such a process are relatively rare. Most of the deep water formation occurs in only a few isolated regions of the world ocean. One important such region is in the Weddell Sea off the coast of Antarctica. In this case a wide continental shelf allows the local water column to be quite shallow and in winter this column can become very cold and dense. This dense water then slides off the continental shelf and flows to the very bottom of the Southern Ocean basin. Such a deep water formation is known to form only in a few areas, such as near the Ross Sea and in the North Atlantic.

Some deep water formations are known to form in the Norwegian Sea and possibly in the Labrador Sea. Once the deep water forms below these areas of the North Atlantic it flows along the bottom western edge of the Atlantic toward the equator almost antiparallel to the Gulf Stream above it.

Some Simple Circulation Models

Middle Latitude Basin Flows. Based upon the principle of wind-driven flows discussed above, one can build a qualitative picture of the flows in a middle latitude ocean basin. Consider the North Atlantic as an example. The Ekman flow in the trade wind region causes a general westward flow north of the equator. The flow is interrupted by the coast of North America (actually the Caribbean Sea and Islands). For these idealized purposes consider the western margin of the North Atlantic to be along a meridian. The irregular middle latitude westerlies cause the flow to move toward the east but with a smaller southerly component because of the Ekman drift. The upshot of this model is a tight boundary current running to the north parallel to the coast on the western margin of the North Atlantic (analog to the Gulf Stream) and breaking away in the middle latitudes to the east. As the flow proceeds east it is returned to the tropics to supply the tropical current. The result is a clockwise flow around the basin featuring a narrow boundary current on the left side of the clock face and a more diffuse return flow to the south on the right side of the clock face.

The Thermohaline Circulation. There is a large-scale circulation driven by density differentials in the world oceans. The deep water formation in the North Atlantic is basically fed by the Gulf Stream running up the western boundary of the North Atlantic. These warm saline waters reach their point of sinking near the coast of Greenland. The Gulf Stream surface current begins with a cross-equatorial flow from the SH first running up the western coast of Africa. Some investigators trace these currents out into an elaborate pattern throughout all hemispheres over the surface waters of the ocean. After the sinking into *deep water* the return flows which are part of this pattern are sometimes called the 'conveyer belt'. We will not pursue this issue since this idealized picture is not universally accepted. But, the part of the conveyer belt operating in the North Atlantic is rather well established. It is part of the overall global density-driven circulation called the thermohaline circulation.

An important question in climate change research is whether the thermohaline circulation has changed or has 'shut down' in the past or whether it might do so in future climate scenarios.

Circumpolar Current. The Southern Ocean surrounds the continent of Antarctica. This ocean is almost completely taken up with a current that circulates clockwise around the polar continent in a coherent form. Of course, it entrains and contributes to waters that feed the deep water formation off the Weddell and Ross Seas.

The Role of Sea Ice. Sea ice forms in the polar oceanic regions. The Arctic Sea covers the North Polar Region and is covered with sea ice – some remaining even through the summer – the so-called perennial sea ice. Sea ice also forms, and its area expands, in winter in the Southern Ocean off the coasts of Antarctica. A similar seasonal oscillation occurs in the North Polar Region. Sea-ice presence and associated growth and decay processes can affect the climate and general circulation of both the atmosphere and the oceans.

Sea ice is far from uniform. Its thickness is highly variable and it is moved around by wind stress as well as by underlying ocean currents. Many cracks and fissures (called leads) occur over the ice surface.

The presence of sea ice affects the reflectivity of sunlight (albedo) impinging on the earth's surface. Sea ice, especially when fresh or snow-covered, is white and very reflective in the visible part of the spectrum. Hence, an expansion of sea-ice area leads to less solar absorption and a cooler atmospheric boundary layer. However, the presence of sea ice cuts off communication of heat and stress between the air above and the sea below. This is complicated by the presence of leads. Climate models must take into account sea-ice dynamics because of these important feedback mechanisms.

Ocean Circulation Models. Ocean circulation models fit for simulation on large computers are now operated routinely. These models have a horizontal resolution of a fraction of 1 degree latitude or longitude. The simulations are becoming very realistic in that many subtle features of the ocean circulation are captured. A major limitation of the ocean model is that many features of the ocean – the analog of the highs and lows in the middle latitudes of the atmosphere – are a factor of 10 smaller. That is to say that the typical eddies which carry much of the heat polewards in the middle latitudes are only about 100 km in size, as opposed to about 1000 km for comparable features in the atmosphere. This means that much more horizontal resolution is necessary, requiring much more computer time. Nonetheless, considerable progress is being made in this field as computers become faster and satellite information providing global coverage becomes more common.

The Coupled Climate System

In the 1990s there were a number of modeling groups merging the atmospheric GCM to the oceanic GCM – these are called coupled oceanic–atmospheric general circulation models (OAGCMs). The work of putting these two media together is not simple, due to the mismatch of temporal and spatial scales. The two also communicate through radiation – the ocean is heated differently in cloud-covered sky than in clear sky. Also, precipitation falls on the ocean freshening its surface waters – likewise for river runoff. Hence, to achieve a satisfactory coupling, many detailed features not necessarily critical for simulating the isolated systems suddenly become important. If care is not exercised the solutions are unstable and 'run away'. But recently there have been reasonably successful attempts to simulate the combined media. The era of coupled ocean–atmosphere modeling lies in the near future and will likely lead to new and startling results.

El Niño and Other Oscillations

One feature of the combined system is the interaction of the global atmospheric circulation with the Tropical Pacific Ocean. This is a cooperative effect in which both partners participate in a nearly equal fashion. The trade winds tend to push ocean surface waters to the west in the Pacific. This can continue for some years, leading to a buildup of a meter or two of sea level in the western part of the ocean. At some point a threshold is reached and the trade winds weaken. This releases the pent-up water in the west and it begins to travel to the east in the form of a slowly moving equatorial wave, taking several months to cross the Pacific Ocean Basin. When the warm waters from the west arrive at the eastern margin of the basin they find cool water.

This cool water (mainly off the coast of Peru) is present because in normal conditions the trade winds (with the help of Ekman drift) push the waters westward from the coast and this brings up cool waters from a few 100 m below. In fact, the cool upwelling waters bring nutrients from below. The upwelling often subsides (but weakly) during the Christmas season. Therefore, the name 'El Niño' was attached to the event, meaning 'the child' in Spanish. But every two to seven years on an irregular basis the wave from the Western Pacific brings the overriding warm water to cover the cooler upwelling for a year or even longer. This is the phenomenon we know as El Niño.

During El Niño the general circulation of the global atmosphere is altered dramatically. The Walker circulation is weakened and the convection that normally occurs over the East Indies comes nearly to a halt because it has now moved to the middle Equatorial Pacific. Drought comes to northern Australia, Borneo, Indonesia, and neighboring lands. Heavy rains now occur in Peru. More subtly, the newly located convection in the area of the dateline releases latent heat in the middle troposphere and this excites long stationary waves stretching from the dateline into North America along a great circle. Along the waves are stationary high- and low-pressure crests and troughs. These nearly stationary features serve to guide storm tracks in North America especially in winter when the westerlies dip well into the US southeast. El Niños typically lead to more rainfall in that region of the world. There are significant effects on rainfall in various other parts of the world as well.

El Niño is only one of several such coupled ocean–atmosphere oscillations currently under study. Climatologists working on such problems are concerned with time-scales of a few months to a few years. Other likely candidates for study include the oscillations which occur in the North and South Atlantic.

Climate Change Simulations

The ultimate use of coupled OAGCMs is the study of climate change brought about by natural variability or induced by external forcings such as the greenhouse effect (*see* **Global warming**). The coupling of the two media are essential for long-term climate change since the time-scales of such change (roughly 100 years) brings the ocean's slow thermal response into play. Some of the more important questions include: How long is the delay provided by the oceans to a forced change in climate? What new feedback mechanisms are in the oceans that include sea-ice cover? How will the frequency and intensity of El Niño respond to climate change? How will the frequency and intensity of tropical storms respond to climate change? How will the thermohaline respond? How will the carbon cycle of the coupled system respond to climate change?

Conclusions

One of the great scientific achievements of the twentieth century has been our increased understanding and practical application of the general circulation of the atmosphere and the ocean. This has happened through the efforts of scientists throughout the world. In fact, much of the achievement would not have occurred without the cooperation of many nations through their generous communication of scientific data and ideas. While we have learned much about the system over the last century, the last section alone tells us that there is much more to be done. As the world's population increases we become more vulnerable to minor weather or oceanic related anomalies, ranging from the incidence of tornadoes to the extinction of **fisheries**. We must press on with our studies to better understand the system as an intellectual enterprise but even more so as a practical guide to a safer world in which to live and prosper.

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(*See also* **Atmospheric dispersion: Basic**; **Global environment monitoring system, water**; **Global environmental change**; **Meteorological extremes**)

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Global ecology

Within the conventions of ecology qua ecology, the notion of global ecology is almost an oxymoron. Traditional ecologists tend to regard ecology as 'The study of the controls over the distribution and abundance of organisms'. According to this criterion, the locus of ecology occurs at a single scale, that of the organism, and research emphasizes organism–environment and organism–organism interactions. Most of the classic questions of both autecology and synecology focus on this scale. Questions regarding the survivorship and growth of individuals in different environments, behavioral ecology, **population dynamics**, predation, and competition, as well as some more modern topics such as symbioses, occur at this single spatial scale.

Global ecology began with Joseph Priestley's experiments in the eighteenth century demonstrating that plants, in effect, purified air and thus allowed animals to survive. These results provided the first hard evidence that the activities of some organisms had impacts on the functioning of other organisms at spatial scales greater than that of the individual. More recently, Redfield [4] demonstrated in the middle of the last century that the structure and behavior of marine organisms played a major role in determining the chemistry of the oceans. In the last 25 years, global ecology has expanded to include a wide range of organismal activities that affect continental and larger spatial scales [3].

Perhaps the highest profile topic of global ecology currently being investigated concerns the exchange of carbon dioxide between the biosphere and the atmosphere (*see* **Meteorology**). At the end of the nineteenth century Svante Arrhenius [1] noted the role of carbon dioxide in maintaining the thermal balance of the Earth's surface. Beginning with the International Geophysical Year in 1958, C.D. Keeling [2] started to measure the average concentration of CO_2 in the lower atmosphere by making repeated measurements at a remote site in Hawaii (see `http://cdiac.eds.ornl.gov/trends/co2/sio-mlo.htm/`). These measurements demonstrated two crucial facts. The first, and most famous, was that average atmospheric CO_2 concentrations were rising exponentially. Later work on gas concentrations in ice cores showed that the beginning of this increase coincided with the start of the Industrial Revolution. Keeling's measurements also demonstrated that the seasons in the northern hemisphere affected the intra-annual changes in CO_2 concentration. During the northern hemisphere's spring and summer CO_2 concentrations dropped, while during the winter they rose. This initial observation spawned entire new research programs (as well as journals) dedicated to investigating the role of the biosphere in regulating atmospheric CO_2 concentrations. Out of this interest has come an increased awareness that the details of microbial and plant physiology, once the province only of physiologists, are essential also for ecological research. Current efforts now examine both the impacts of biological activity on the atmosphere and the effects of CO_2 concentration on plant growth,