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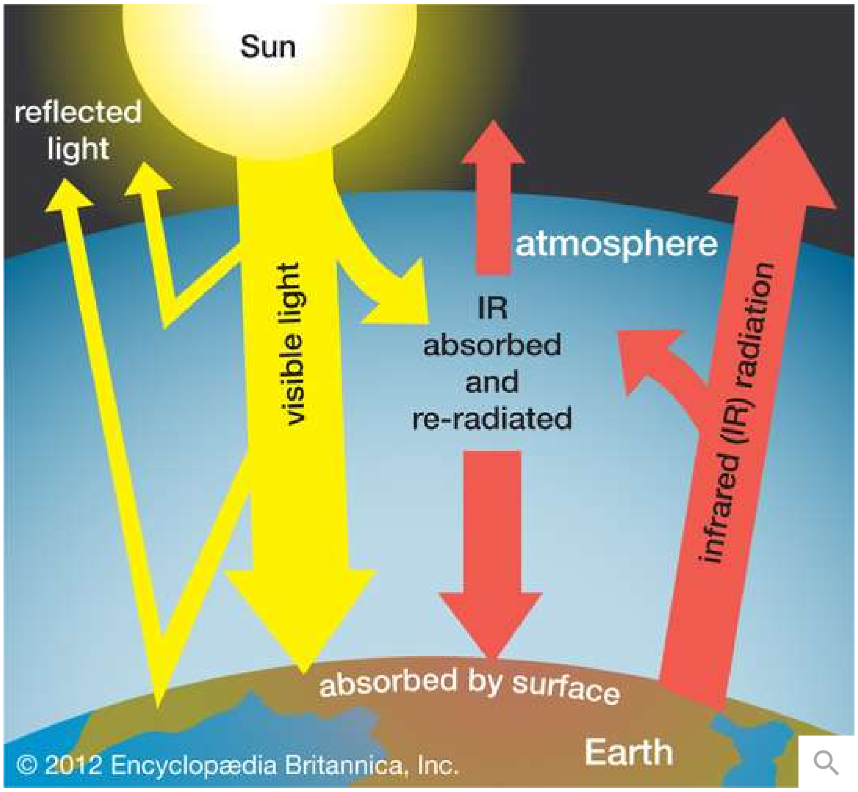
# The role of the biosphere in the Earth-atmosphere system

## The biosphere and Earth's energy budget

Biogenic gases in the atmosphere play a role in the dynamics of Earth's planetary radiation budget, the thermodynamics of the planet's moist atmosphere, and, indirectly, the mechanics of the fluid flows that are Earth's planetary wind systems. In addition, human cultural and economic activities add a new dimension to the relationship between the biosphere and the atmosphere. While humans are biologically trivial compared with bacteria in the exchange of gases with the atmosphere, chemical compounds produced from human industrial activities and other economic enterprises are changing the gaseous composition of the atmosphere in climatically significant ways. The largest changes involve the harvesting of ancient carbon stores. This organic material has been transformed into fossil fuels (coal, petroleum, natural gas, and others) by geologic processes acting upon the remains of plants and animals over many millions of years. Different forms of carbon may be burned and thus used as energy sources. In so doing, organic carbon is converted into carbon dioxide. Additionally, humans are also burning trees, grasses, and other biomass for cooking purposes and clearing the land for agriculture and other activities. The combination of burning both fossil fuels and biomass is enriching the atmosphere with carbon dioxide and adding to the essential reservoir of greenhouse gases (*see* global warming).




**greenhouse effect on Earth**

The greenhouse effect on Earth. Some incoming sunlight is reflected by Earth's atmosphere and surface, but most is absorbed by the surface, which is warmed. Infrared (IR) radiation is then emitted from the surface. Some IR radiation escapes to space, but some is absorbed by the atmosphere's greenhouse gases (especially water vapour, carbon dioxide, and methane) and reradiated in all directions, some to space and some back toward the surface, where it further warms the surface and the lower atmosphere.

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Earth's atmosphere is largely transparent to sunlight. Of the sunlight absorbed by the entire Earth-atmosphere system, about one-third is absorbed by the atmosphere and two-thirds by Earth's surface. Sunlight is absorbed by the molecules of the atmosphere, by cloud droplets, and by dust and debris. Though oxygen and nitrogen make up nearly 99 percent of the atmosphere, these diatomic molecules do not vibrate in a way that permits them to absorb terrestrial radiation. They are largely transparent to outgoing terrestrial radiation as well as to incoming solar radiation.

Over the continents, the surface cover of vegetation is the principal absorbing medium of Earth's surface, although other surfaces such as bare rock, sand, and water also absorb solar radiation. At night, absorption at the surface (that is, below 1.2 metres [4 feet]) is reradiated, in the form of long-wave infrared radiation, away from Earth's surface back toward space. Most of this infrared radiation is absorbed by the principal biogenic trace gases of the atmosphere—the so-called greenhouse gases: water vapour, carbon dioxide, and methane. Without these biogenic greenhouse gases, Earth would be 33 °C (59 °F) colder on average than it is. A moderate-emission scenario from the 2007 Intergovernmental Panel on Climate Change (IPCC) report predicts that the continued addition of greenhouse gases from fossil fuels will increase the average global temperature by between 2.3 and 4.3 °C (4.1 and 7.7 °F) over the next century. Other scenarios, predicting greater greenhouse gas emissions, forecast even greater global warming.



## The cycling of biogenic atmospheric gases

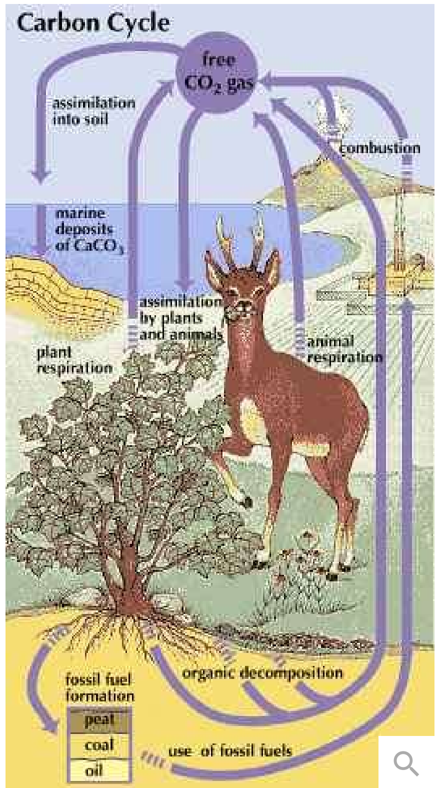
The cycling of oxygen, nitrogen, water vapour, and carbon dioxide, as well as the trace gases—methane, ammonia, various oxides of nitrogen and sulfur, and non-methane hydrocarbons—between the atmosphere and the biosphere results in relatively constant proportions of these compounds in the atmosphere over time. Without the continuous generation of these gases by the biosphere, they would quickly disappear from the atmosphere.

**Average composition of the atmosphere**

| gas | composition by volume (ppm)* | composition by weight (ppm)* | total mass ($10^{20}$ g) |
|---|---|---|---|
| **nitrogen** | 780,900 | 755,100 | 38.648 |
| **oxygen** | 209,500 | 231,500 | 11.841 |
| **argon** | 9,300 | 12,800 | 0.655 |
| **carbon dioxide** | 386 | 591 | 0.0299 |
| **neon** | 18 | 12.5 | 0.000636 |
| **helium** | 5.2 | 0.72 | 0.000037 |
| **methane** | 1.5 | 0.94 | 0.000043 |
| **krypton** | 1.0 | 2.9 | 0.000146 |
| **nitrous oxide** | 0.5 | 0.8 | 0.000040 |
| **hydrogen** | 0.5 | 0.035 | 0.000002 |
| **ozone**** | 0.4 | 0.7 | 0.000035 |
| **xenon** | 0.08 | 0.36 | 0.000018 |

*ppm = parts per million.
**Variable, increases with height.

The carbon cycle, as it relates to the biosphere, is simple in its essence. Inorganic carbon (carbon dioxide) is converted to organic carbon (the molecules of life). To complete the cycle, organic carbon is then converted back to inorganic carbon. Ultimately, the carbon cycle is powered by sunlight as green plants and cyanobacteria (blue-green algae) use sunlight to split water into oxygen and hydrogen and to fix carbon dioxide into organic carbon. Carbon dioxide is removed from the atmosphere, and oxygen is added. Animals engage in aerobic respiration, in which oxygen is consumed and organic carbon is oxidized to manufacture inorganic carbon dioxide. It should be noted that chemosynthetic bacteria, which are found in deep-ocean and cave ecosystems, also fix carbon dioxide and produce organic carbon. Instead of using sunlight as an energy source, these bacteria rely on the oxidation of either ammonia or sulfur.




The carbon cycle is the complex path that carbon follows through the atmosphere, oceans, soil, and plants and animals.

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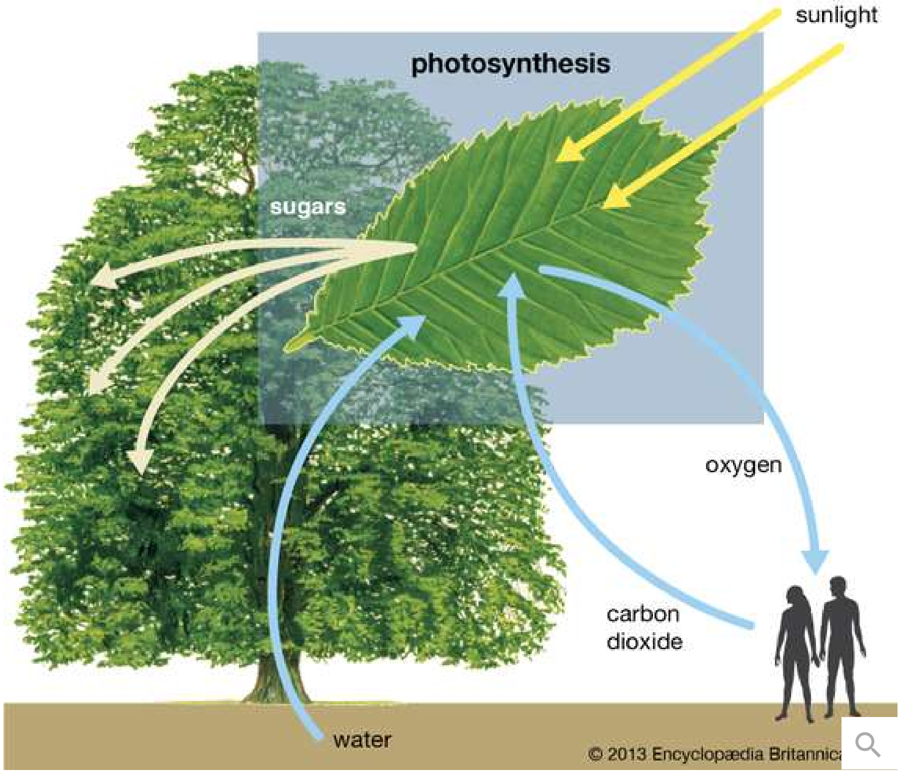
The carbon cycle is fully coupled to the oxygen cycle. Each year, photosynthesis fixes carbon dioxide and releases 100,000 megatons of oxygen to the atmosphere. Respiration by animals and living organisms consumes about the same amount of oxygen and produces carbon dioxide in return. Oxygen and carbon dioxide are thus coupled in two linked cycles. On a seasonal basis, an enrichment of atmospheric carbon dioxide occurs in the winter half of the year, whereas a drawdown of atmospheric carbon dioxide takes place during the summer.




Green plants such as trees use carbon dioxide, sunlight, and water to create sugars. Sugars provide the energy that makes plants grow. The process creates oxygen, which people and other animals breathe.

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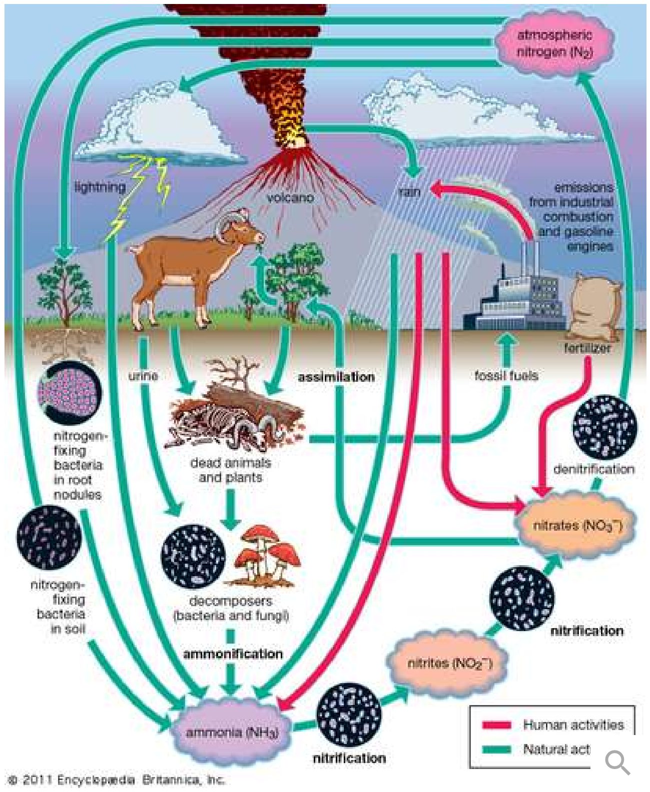
The nitrogen cycle begins with the fixing of inorganic atmospheric nitrogen ($N_2$) into organic compounds. These nitrogen-containing compounds are used by organisms and, through the process of denitrification, are converted back to inorganic atmospheric nitrogen. Ammonia and ammonium ions are the products of nitrogen fixation and may be incorporated into some organisms as organic nitrogen-containing molecules. In addition, ammonium ions may be oxidized to form nitrites, which can be further oxidized by nitrifying bacteria into nitrates. Though both nitrites and nitrates may be used to make organic nitrogen-containing molecules, nitrates are especially useful for plant growth and are key compounds that support both terrestrial and aquatic food chains. Nitrates may also be denitrified by bacteria to produce nitrogen gas. This process completes the nitrogen cycle.




**nitrogen cycle**

The nitrogen cycle transforms diatomic nitrogen gas into ammonium, nitrate, and nitrite compounds.

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The nitrogen cycle is coupled to both the carbon and oxygen cycles. Seventy-eight percent of the gases of the atmosphere by volume is diatomic nitrogen. Diatomic nitrogen is the most stable of the nitrogen-containing gases of the atmosphere. Only 300 megatons of nitrogen must be produced each year by denitrifying bacteria to account for losses. These losses mostly occur during lightning discharges and during nitrogen-fixation activities by blue-green algae and nitrogen-fixing bacteria. (The latter are found in the root nodules of certain plants called legumes.) Nitrogen-fixing bacteria use atmospheric nitrogen to produce oxides of nitrogen. Denitrifying bacteria convert nitrates in soils and wetlands to nitrogen gas, which is then returned to the atmosphere.

Nitrous oxide occurs in trace amounts (0.3 ppm) in the atmosphere. Between 100 and 300 megatons of nitrous oxide are produced by soil and marine bacteria each year to maintain this concentration. In the atmosphere, nitrous oxide is short-lived because it is quickly broken down by ultraviolet light. Nitric oxide (NO), a minor contributor in the breakdown of stratospheric ozone, is the by-product of this reaction.

Like nitrous oxide, ammonia is also produced in soils and marine waters by bacteria and escapes into the atmosphere. Nearly 1,000 megatons are added to the atmosphere each year. Ammonia decreases the acidity of precipitation and serves as a nutrient for plants when returned to the land via precipitation.

There are six major sulfur-containing biogenic atmospheric gases that are part of the sulfur cycle. They include hydrogen sulfide, carbon disulfide ($CS_2$), carbonyl sulfide (COS), dimethyl sulfide (DMS; $C_2H_6S$), dimethyl disulfide ($[CH_3S]_2$), and methyl mercaptan ($CH_3SH$). Sulfur dioxide ($SO_2$) is an oxidation product of these sulfur gases, and it is also added to the atmosphere by volcanoes, burning biomass, and anthropogenic sources (i.e., smelting metals and coal ignition). $SO_2$ is removed from the atmosphere and returned to the biosphere in rainfall. The increased acidity of rain and snow from anthropogenic additions of $SO_2$ and oxides of nitrogen is often referred to as "acid precipitation." The acidity of this precipitation and other phenomena, such as "acid fog," is partly cancelled by the release of ammonia in the atmosphere.

The concentration of methane at any one time in the atmosphere is only about 1.7 ppm. Though only a trace gas, it is highly reactive and plays a key role in the chemical reactions that control the composition of the atmosphere. Methanogenic bacteria in wetland sediments decompose organic matter and release 1,000 megatons of gaseous methane to the atmosphere per year. In the lower atmosphere, methane reacts with oxygen to produce water and carbon dioxide. Each year 2,000 megatons of oxygen are removed from the atmosphere by this mechanism. This loss of oxygen must be replaced by photosynthesis. Some methane reaches the upper stratosphere, where its interaction with oxygen is a major source of upper stratospheric moisture. Within wetlands, bacteria produce methyl halide compounds (methyl chloride [$CH_3Cl$] and methyl iodide [$CH_3I$] gases), whereas these same methyl halides are produced in forests by fungi. These gases, upon reaching the stratosphere, regulate the production of stratospheric ozone by contributing to its natural breakdown (*see* ozonosphere). Without the continual production of methane by methanogenic bacteria, the oxygen concentration of the atmosphere would increase by 1 percent in only 12,000 years. Dangerously high levels of oxygen in the atmosphere would greatly increase the incidence of wildfires. If the oxygen concentration of Earth's atmosphere rose from its current concentration of 21 percent to 25 percent, even damp twigs and grass would easily ignite. Non-methane hydrocarbons of terrestrial origin are generally well mixed in the free atmosphere above the planetary boundary layer (PBL; *see below*). These organic particles weaken incoming solar radiation as it passes through the atmosphere, and reductions of 1 percent have been recorded.

## Biosphere controls on the structure of the atmosphere

Because the biosphere plays a key role in the flux of energy from the surface to the atmosphere, it also contributes to the structure of the atmosphere. Three major fluxes are important: the direct transfer of heat from the surface to the atmosphere by conduction and convection (sensible heating), the energy flux to the atmosphere carried by water vapour via evaporation and transpiration from the surface (latent heat energy), and the flux of radiant energy from the surface to the atmosphere (infrared terrestrial radiation). These fluxes differ in the altitude at which the heating of the air takes place and thus contribute to the thermal structuring of the atmosphere. Sensible heating primarily warms the planetary boundary layer (PBL) of the atmosphere. In marine areas, the PBL occurs in the lowest 1 km (3,300 feet); in heavily vegetated areas, the PBL occurs in the lowest 1 to 2 km (3,300 to 6,600 feet); and in arid regions, it occurs in the lowest 4 or 5 km (13,100 to 16,400 feet) of the atmosphere. In contrast, the latent heat of the atmosphere is released when the water vapour is converted into cloud droplets by condensation. Heating by latent energy release generally occurs above the PBL.

On the other hand, heating of the atmosphere by radiation from the surface depends on the density of the atmosphere and its water vapour content. Radiative heating from the surface declines with increasing altitude. The availability of water to evaporate from the surface limits the sensible heating of the air near the surface and so limits the maximum daytime surface air temperature (*see below*).

## Biosphere controls on the planetary boundary layer

The top of the planetary boundary layer (PBL) can be visually marked by the elevation of the base of the clouds. In addition, the PBL can also be denoted by a thin layer of haze often seen by passengers aboard airplanes during takeoff from airports. During the day, the air within the PBL is thoroughly mixed by convection induced by the heating of Earth's surface. The thickness of the PBL depends on the intensity of this surface heating and the amount of water evaporated into the air from the biosphere. In general, the greater the heating of the surface, the deeper the PBL. Over deserts, the PBL may extend up to 4 or 5 km (13,100 or 16,400 feet) in altitude. In contrast, the PBL is less than 1 km (0.6 mile) thick over ocean areas, since little surface heating takes place there because of the vertical mixing of water. The wetter the air advected into the region and the greater the additional water added by evaporation and transpiration, the lower the height of the top of the PBL. For every 1 °C (1.8 °F) increase in daily maximum surface temperature for a well-mixed PBL, the top of the PBL is elevated 100 metres (about 325 feet). In New England forests during the days following the spring leafing, it has been shown that the top of the PBL is lowered to between 200 and 400 metres (650 and 1,300 feet). By contrast, during the months before the leafing out, the PBL thickens from solar heating as the sun rises higher in the sky and day length increases.

If convective mixing of the air in the PBL is vigorous, convection currents may penetrate through the temperature inversion at the top of the PBL. The cooling of the lifting air initiates the condensation of water vapour and the development of miniscule particles of liquid water called cloud droplets. The small clouds just above the PBL are known as planetary boundary layer clouds. These clouds scatter direct sunlight. As the ratio of diffuse sunlight to direct beam sunlight increases, greater levels of photosynthetic productivity are favoured in the biosphere below. The result is a dynamic synergy between the atmosphere and biosphere.

The landscapes of most human-dominated ecosystems are decidedly “patchy” in their geography (*see below*). Cities, suburbs, fields, forests, lakes, and shopping centres both heat and evaporate water into the air of the PBL according to the nature of the surfaces involved. Convection and the prospect of breaking through the top of the PBL vary markedly across such heterogeneous landscapes. These upward and downward currents or vertical eddies within the PBL transfer mass and energy upward from the surface. The frequency, timing, and strength of convective weather elements, including thunderstorms, vary according to the patchiness of the land use and land cover pattern of the area. In general, the greater the patchiness of the landscape and the earlier the hour in the day, the more frequent and more intense these rain-producing systems become.

In the absence of an organized storm in the region, the air above the PBL sinks gently and the air below lifts. At the top of the PBL, a small inversion, where temperatures increase with height, develops. This inversion essentially becomes a stable layer in the atmosphere. Emissions from the biosphere below are thus contained within the PBL and may build up below this layer over time. Consequently, the PBL may become quite turbid, hazy, or filled with smog.


**Learn how pollution from industrial emissions and car exhaust is trapped against mountains to cause smog**
Smog formation and entrapment over Los Angeles.

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When the sinking from above is vigorous, the PBL inversion grows in thickness. This situation has the effect of hindering the development of thunderstorms, which depend on rapidly rising air. This often occurs over southern California, and thus the chance of thunderstorms forming there is small. Emissions from both the biosphere and from anthropogenic activities accumulate in this part of the atmosphere, and pollution may build up to such an extent that health warnings may be required. In locations free of temperature inversions, convection processes are strong enough, particularly during the summer months, that emissions are scavenged and quickly lifted by thunderstorms to regions high above the PBL. Often, acidic compounds from these emissions are returned to the surface in the precipitation that falls.



## Biosphere controls on maximum temperatures by evaporation and transpiration

Solar radiation is converted to sensible and latent heat at Earth's surface. A change in sensible heat results in a change in the temperature of a medium, whereas energy stored as latent heat is used to drive a process, such as a phase change in a substance from its liquid to its gaseous state, and does not produce a change in temperature. Thus, the daily maximum surface temperature at a given location is dependent on the amount of radiant energy converted to sensible heat. Water available for evaporation increases latent heating by adding water vapour to the atmosphere. As a result, relatively little energy remains to heat the air, and thus the sensible heating of the air near the ground is minimized. In addition, daily maximum temperatures are not as high in locations with strong latent heating.

As day length increases from winter to summer, sensible heating and maximum surface temperatures rise. In the U.S. Midwest, prior to the leafing out of vegetation in the springtime and the resulting rise in evaporation and transpiration, sensible heating causes an average increase in maximum surface temperatures of only about 0.3 °C (0.5 °F) per day. The process of leaf production creates a surge in evaporation and transpiration and results in increased latent heating and reduced sensible heating. After leafing, since most of the available thermal energy is used to convert liquid water to water vapour rather than to heat the air, the average day-to-day rise in daily maximum temperatures is reduced to about 0.1 °C (0.2 °F) per day.

This effect extends upward through the atmosphere. Prior to leafing out, the one-kilometre-thick layer occurring between the 850-to-750-millibar pressure level (which typically occurs between 1,650 and 2,750 metres [5,400 and 9,000 feet]) in the Midwest warmed at the rate of 0.1 °C (0.2 °F) per day. Following leafing out, the warming rate fell to 0.02 °C (0.04 °F) per day. Scientists have used computer models of the atmosphere to study the effect of transpiration from vegetation on maximum surface air temperatures. In these models, the variable controlling transpiration by vegetation was "turned off," and the character of the resulting modeled climate was studied. By subtracting the effect of transpiration, temperatures in central North America and on the other continents were predicted to equilibrate at a very hot 45 °C (113 °F). Such warming is nearly realized in desert areas where moisture is unavailable for transpiration.

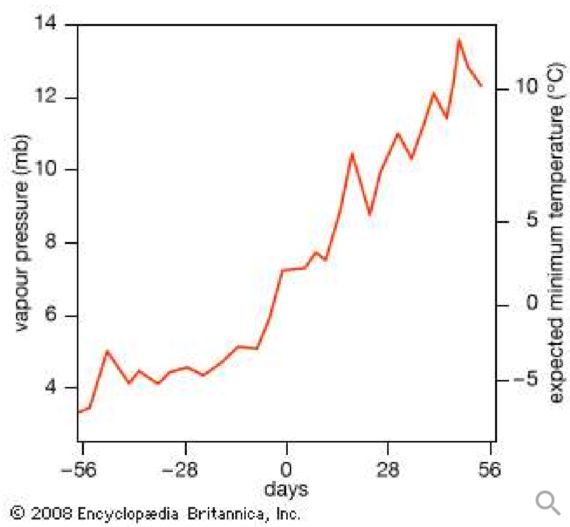
## Biosphere controls on minimum temperatures

During the late 1860s, British experimental physicist John Tyndall, based on his studies of the infrared radiation absorption by atmospheric gases, concluded that nighttime minimum temperatures were dependent on the concentration of trace gases in the atmosphere. Of these gases, water vapour had the greatest impact. To emphasize the significance of water vapour on decreases in air temperature during the night, he wrote that if all the water vapour in the air over England was removed even for a single night, it would be "attended by the destruction of every plant which a freezing temperature could kill." As a result, it follows that the greater the water content of the atmosphere, the lower the radiative loss of energy to the sky and the less the surface atmosphere is cooled. Thus, locations with substantial amounts of water vapour experience reduced nocturnal cooling.

Water vapour in the atmosphere also limits the extent to which temperatures fall at night. This limiting temperature is known as the dew point, which is defined as the temperature at which condensation begins. Over North America east of the 100th meridian (a line of longitude traditionally dividing the moist eastern part of North America from drier western areas), average nighttime minimum temperatures are within a degree or two of the dew point temperature. Upon nocturnal cooling, the dew point is reached, condensation begins, and latent energy is converted to heat. Additional temperature falls are retarded by this release of heat to the atmosphere. A significant fraction of the water in the atmosphere over the continents comes from the evaporation of water from soils and the transpiration from vegetation. Transpired water directly moderates temperature by increasing humidity and thus raising the dew point. As a consequence, the amount of outgoing terrestrial radiation released to space is reduced. This results in the elevation of the minimum temperature of the air above what it would otherwise be.

Dew often forms on grass during cool nights.

The effect of spring leafing on the buildup of humidity in the lower atmosphere has received the attention of researchers in recent years. In the late 1980s, American climatologists M.D. Schwartz and T.R. Karl used the superimposed epoch method to study the climate before and after the leafing out of lilac plants in the spring in the U.S. Midwest. (This method uses time series data from multiple locations, which can be compared to one another by adjusting each data set around the respective onset date of lilac blooming.) In the illustration, the *x*-axis marks days before and after leafing, whereas the *y*-axis shows the related changes in the vapour pressure of the atmosphere. A second *y*-axis follows the day-by-day changes in minimum temperatures. Prior to the average date of leafing, the atmospheric humidity (vapour pressure) is relatively constant and minimum temperatures hover near freezing. At leafing, there is an abrupt increase in atmospheric humidity. Following leafing, daily minimum temperatures also increase abruptly. Although frosts are possible until June 10 in many parts of the Midwest, the chances of frost decline as the atmosphere is humidified.




**lilac bud break, humidity, and temperature**

Graph of atmospheric vapour pressure and expected minimum temperature for 56 days prior to and following the average day of the leafing out of lilac plants, based on data compiled by M.D. Schwartz and T.R. Karl, 1990.

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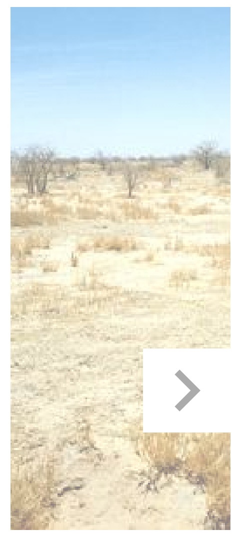
## Climate and changes in the albedo of the surface

The amount of solar energy available at the surface for sensible and latent heating of the atmosphere depends on the albedo, or the reflectivity, of the surface. Surface albedos vary by location, season, and land cover type. The albedo of unvegetated ground devoid of snow ranges from 0.1 to 0.6 (10 to 60 percent), while the albedo of fully forested lands ranges from 0.08 to 0.15. An increase of 0.1 in regional albedo has been associated with a 20 percent decline in rainfall events connected with thunderstorms. Equivalent reductions in both evaporation and transpiration have also been reported in areas with sudden increases in albedo.

The greatest changes in albedo occur in regions undergoing desertification and deforestation. Depending on the albedo of the underlying soil, reductions in vegetative land cover may give rise to albedo increases of as much as 0.2. Model studies of the vegetative zone known as the Sahel in Africa reveal that albedo increased from 0.14 to 0.35 due to desertification occurring during the 20th century. This coincided with a 40 percent decrease in rainfall. In addition, it is likely that the clearing of forests and prairies for agricultural crops over the past several hundred years has altered the albedo of extensive regions of the middle latitudes.



**deforestation**

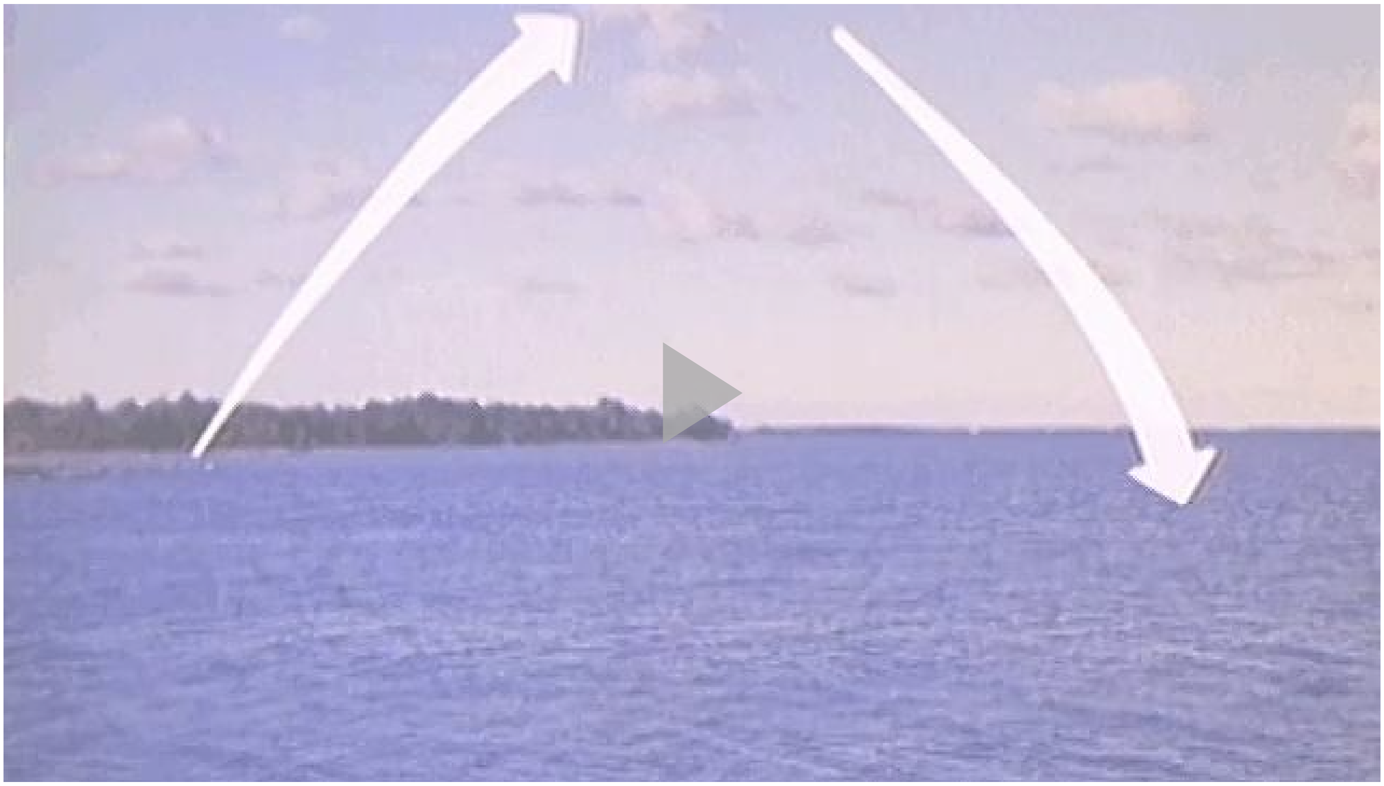
Smoldering remains of a plot of deforested land in the Amazon Rainforest of Brazil. Annually, it is estimated that net global deforestation accounts for about two gigatons of carbon emissions to the atmosphere.

Contemporary agricultural practices give rise to large variations in albedo from season to season as the land passes through the cycle of tilling, planting, crop growth, and harvest. At larger scales, an agricultural mosaic often emerges as each different plot of ground is covered by plantings of a single species. Viewed from the air, landscapes in the middle latitudes appear as a heterogeneous mix of forests, grasslands, meadows, water bodies, farmlands, wetlands, and urban types. The resultant patchiness in the landscape produces a patchiness in surface albedo. The mosaic of land use types creates a mix in the fluxes of sensible and latent heat to the atmosphere. Such changes to the heat flux have been shown to cause changes in the timing, intensity, and frequency of summer thunderstorms.

## The effect of vegetation patchiness on mesoscale climates

The establishment of vegetation bands or patches 50 to 100 km (30 to 60 miles) in width in semiarid regions could increase atmospheric convection and precipitation beyond that expected over areas of uniform vegetation. This convection creates spatial differences in the upward and downward wind velocities and contributes to the development of mesoscale (20 to 200 km [12 to 120 miles]) circulation in the atmosphere (*see* Upper-level winds: Characteristics). For example, when creating models for forecasting atmospheric conditions on the Great Plains and along the Front Range of the Rocky Mountains, the mix of land cover and vegetation types must be specified to properly relate the fluxes of momentum and sensible and latent heat to the larger-scale circulation of the atmosphere. Proper calculations are also necessary to estimate rainfall. In addition, the specific location and hour of the day that thunderstorms occur depend on the heterogeneity of the vegetation cover of this region. Field observations have shown that the heterogeneity of surface roughness (small-scale irregularities in topography), soil moisture, forest coverage, and transpiration affect the location and pace of the formation of convective clouds and rainfall. Both convection and thunderstorm development tend to occur earlier in the day in heterogeneous landscapes.

## Biosphere controls on surface friction and localized winds



**Learn how mountains, bodies of water, and human habitation affect atmosphere activity**
Factors affecting the localized movement of air.

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Averaged annually over Earth's entire surface, the Sun provides about 345 watts per square metre of energy. About 30 percent of this energy is reflected away to space and is never used in the Earth-atmosphere system. Of that which remains, a little less than 1 percent (3.1 watts per square metre) accelerates the air by generating winds. An equal amount of energy must eventually be lost, or else wind speeds would perpetually increase.

Earth as a thermodynamic system is dissipative—the mechanical energy of the winds is eventually converted to heat through friction. Over the continents, it is the combination of terrain and the veneer of vegetation that offers the frictional roughness to dissipate the surface winds and convert this kinetic energy into heat. Marine winds approaching the British Isles average about 12 metres per second (27 miles per hour), but they are decelerated to 6 metres per second (13 miles per hour) because of the friction of the landscape's surface shortly after the winds make landfall. Without vegetation cover, the continents would offer much less friction to the wind, and wind speeds in unvegetated landscapes would be nearly twice as fast as those in vegetated landscapes.

The correct specification of Earth's surface roughness due to vegetation, for use in computer models of the atmosphere, is critical to proper model performance. If the height of the terrain and vegetation are not specified correctly, the patterns of Earth's winds, global geography, and rainfall will be poorly modeled. When modeling newly desertified areas, such as the Sahel, it is important to understand that desertification creates vegetation of lower stature and thus lower surface roughness values. As a result, both wind velocities and wind direction could change from previous patterns over landscapes with taller vegetation.

The extent of this impact of the biosphere on the atmosphere is revealed in climate model studies. One such study modeled the influence of reduced vegetation on surface roughness over the Indian subcontinent and provided evidence for a weaker monsoon and reduced rainfall. Given that much of the northwest third of India underwent a severe desertification and cultural collapse near the beginning of historical times, the role cultures play in vegetation reduction and climate change should not be ignored.

The vegetation cover of the continents is not passive in response to the winds. Greenhouse-grown trees subjected to mechanical forces designed to mimic the winds lay down new woody tissue called "reaction wood," which results in a stiffer tree over time. This material helps trees become more resilient and offer more frictional resistance to wind. This negative feedback, where increased winds result in stiffer vegetation and thereby subsequently reduced wind speeds, might well apply at the global scale by balancing the energy used to heat and accelerate the air (3.1 watts per square metre) with the surface friction needed to dissipate it.



## Biosphere impacts on precipitation processes

### Cloud condensation nuclei

The formation and subsequent freezing of cloud droplets depend on the presence of cloud condensation nuclei and ice nuclei, respectively. Significantly, the biosphere is a major source of both of these kinds of nuclei. Over the continents, condensation nuclei are readily available and are of biogenic as well as anthropogenic origin. Examples of condensation nuclei include sea salt, small soil particles, and dust.

As atmospheric convection increases with the heating of the day, cloud condensation nuclei are mixed into and above the planetary boundary layer and into the troposphere. In the bottom 0.5 km (the lowest 1,600 feet or so) of the atmosphere, nuclei typically number 2.2 × $10^{10}$ per cubic metre. In the next 0.5 km (between 1,600 and 3,300 feet) above, half as many nuclei are found. The number of condensation nuclei continues to decline with increased altitude. Furthermore, in general, the number of nuclei in the air over land is 10 times higher than over the oceans.

Cloud condensation nuclei are generally abundant. They do not limit cloud formation over the continents; however, low numbers of condensation nuclei over the oceans may limit cloud formation there. In addition to natural sources, particulates from fuel combustion and sulfur dioxide gas resulting from high sulfur fuels also contribute to the load of condensation nuclei over the continents. Both the number and kind of condensation nuclei present in the atmosphere affect the cloudiness and the brightness of clouds in a given region. In this way, condensation nuclei play a significant role in determining both regional and global albedo.

There is a type of condensation nuclei that forms in the marine air over the margins of continents. Though these nuclei are often few in number, they play a large role in cloud formation near the coasts of continents and may contribute significantly to both planetary albedo and global average temperature. Typically, sources of condensation nuclei in marine air are sulfate aerosols formed from the biogenic production of dimethyl sulfide (DMS) by marine algae. Given that DMS production increases with sea surface temperatures, a negative feedback may result. The central idea in this feedback hypothesis is that warmer waters result in the increased production of condensation nuclei by phytoplankton and thus produce more clouds. Increased cloudiness shades the ocean surface and results in lower temperatures that limit condensation nuclei production. It is estimated that a 30 percent increase in marine condensation nuclei would increase planetary albedo by 0.005 (0.5 percent) or produce a 0.7 percent reduction in solar radiation and a planetary average temperature decrease of 1.3 °C (2.3 °F). The sensitivity of this negative feedback on planetary temperatures remains in active debate.

### Biogenic ice nuclei

As water vapour condenses onto condensation nuclei, the droplets grow in size. Growth proceeds at relative humidity as low as 70 percent, but the rate of growth is very slow. Growth by condensation is most rapid where the air is slightly supersaturated with water vapour. At this point, cloud droplets typical of the size of fog droplets arise. Should temperatures fall to the level where freezing begins, the temperature difference between the droplet and the surrounding air (the vapour pressure deficit) strongly favours rapid condensation into the crystalline lattice of an ice particle.

Ice particles that grow rapidly soon reach sizes where they begin to fall. As they fall, they collide and merge with smaller droplets and thereby grow larger.

The formation of ice is of critical importance. A droplet of pure water, such as distilled water, will automatically freeze in the atmosphere at a temperature of −40 °C (−40 °F). Freezing at warmer temperatures requires a substance upon which ice crystallization can take place. The common clay mineral kaolinite, a contaminant of the droplet, raises this freezing point to around −25 °C (−13 °F). Furthermore, silver iodide, often used in cloud seeding to encourage rainfall, and sea salts also cause ice to form at −25 °C. Freezing at still warmer temperatures is most common with biogenic ice nuclei. Upon ice formation, heat energy on the order of 80 calories per gram of water frozen are released. This energy increases the sensible heat of the air and causes the air to become more buoyant. The process of ice formation encourages convection, cloudiness, and precipitation from clouds.

The decomposition of organic matter is a major source of biogenic ice nuclei. Ice crystal formation has been shown to occur at temperatures as warm as −2 to −3 °C (28.5 to 26.6 °F) when biogenic ice nuclei are involved. The common freezing temperature for biogenic nuclei varies systematically according to biome and latitude. The coldest freezing-temperature nuclei occur above the tropics, whereas the warmest occur above the Arctic. Freezing produces greater buoyancy of the particles and helps them to reach higher vertical velocities within the clouds. The vertical motions and the larger droplet size that occur with biogenic materials favour the charge separation needed to produce lightning. Subsequently, oceanic areas with few biogenic ice nuclei are also areas of low lightning frequency. The production of biogenic nuclei from organic matter decomposition is greatest during the warm months when bacterial decomposition is greatest.

### Recycled rainfall

The water that is transpired into the atmosphere from the biosphere is eventually returned to the surface as precipitation. This vegetation-transpiration component of the hydrologic cycle is referred to as “recycled rainfall.” While the oceans are the major source of atmospheric water vapour and rainfall, water from plant transpiration is also significant. For example, in the 1970s and ’80s, analyses performed by American meteorologist Michael Garstang on the city of Manaus, Brazil, in the Amazon basin revealed that around 20 percent of the precipitation came from water transpired by vegetation; the remaining 80 percent of this precipitation (an estimate made by German American meteorologist Heinz Lettau in the 1970s) was generated by the Atlantic Ocean. Isotopic studies of rainwater collected at various points in the Amazon basin indicated that nearly half of the total rain came from water originating in the ocean and half transpired through the vegetation. Evidence of the proportion of transpired water in rainfall reaching as high as 88 percent has been reported for the Amazon foothills of the Andes. General climate circulation models indicate that, without transpired water from plants, rainfall in the central regions of the continents would be greatly reduced. As a general rule, the farther the distance from oceanic water sources, the higher the fraction of rainwater originating from transpiration.