



Abiogenesis

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Abiogenesis (British English: /ˌeɪ.baɪoʊˈdʒɛnɪsɪs, -ˌbaɪə-, /-biːoʊ-, -biːə-^{[1][2][3][4]}), **biopoiesis**,^[5] \by-o-po-ee-sis  or informally, the **origin of life**,^{[6][7][8]} is the natural process by which life arises from non-living matter, such as simple organic compounds.^{[6][7][9][10]} Abiogenesis is studied through a combination of paleontology, laboratory experiments and extrapolation from the characteristics of modern organisms, and aims to determine how pre-life chemical reactions gave rise to life on Earth.^[11]

The study of abiogenesis can be geophysical, chemical, or biological,^[12] with more recent approaches attempting a synthesis of all three.^[13] Life itself is dependent upon the specialized chemistry of carbon and water and is largely based upon five different families of chemicals. Lipids are fatty molecules comprising large chemical chains of hydrocarbons and play an important role in the structure of living cell membranes, actively and passively determining the transport of other molecules into and out of cells. Carbohydrates are sugars, and as monomer units can be assembled into polymers called polysaccharides, such as cellulose, the rigid chemical of most plant cell walls. Nitrogenous bases are organic molecules in which the amine group of nitrogen, combined with two hydrogen atoms, plays an important part. Chlorophyll is based upon a porphyrin ring derived from amine monomer units, and is important in the capture of the energy needed for life. Nucleic acid monomers are made from a carbohydrate monosaccharide a nitrogenous base and one or more high energy phosphate groups. When joined together they form either the unit of inheritance, the gene, made from DNA or RNA, which translates the genetic information into protein structures. The monomer unit of a protein is usually one of 20 amino acids, comprising an amine group, a hydrocarbon, and a carboxylic acid. Through a condensation reaction, in which the carboxylic acid of one amino acid is linked to the amine of another with removal of a water molecule, a peptide bond is formed. Polymers of amino acids are termed proteins and these molecules provide many catalytic metabolic functions for living processes. Any successful theory of abiogenesis must explain the origins and interactions of these five classes of molecules.^[14]

Many approaches to abiogenesis investigate how self-replicating molecules, or their components, came into existence. It is generally thought that current life on Earth is descended from an RNA world,^[15] although RNA-based life may not have been the first life to have existed.^{[16][17]} The classic Miller–Urey experiment and similar research demonstrated that most amino acids, the basic chemical constituents of the proteins used in all living organisms, can be synthesized from inorganic compounds under conditions intended to replicate those of the early Earth. Various external sources of energy that may have triggered these reactions have been proposed, including lightning and radiation. Other approaches ("metabolism-first" hypotheses) focus on understanding how catalysis in chemical systems on the early Earth might have provided the precursor molecules necessary for self-replication.^[18] Complex organic molecules have been found in the Solar System and in interstellar space, and these molecules may have provided starting material for the development of life on Earth.^{[19][20][21][22]}

The panspermia hypothesis alternatively suggests that microscopic life was distributed to the early Earth by meteoroids, asteroids and other small Solar System bodies and that life may exist throughout the Universe.^[23] It is speculated that the biochemistry of life may have begun shortly after the Big Bang, 13.8 billion years ago, during a habitable epoch when the age of the universe was only 10 to 17 million years.^{[24][25]} The panspermia hypothesis proposes that life originated outside the Earth, not how life came to be.



Precambrian stromatolites in the Siyeh Formation, Glacier National Park. In 2002, a paper in the scientific journal *Nature* suggested that these 3.5 Ga (billion years) old geological formations contain fossilized cyanobacteria microbes. This suggests they are evidence of one of the earliest life forms on Earth.

Nonetheless, Earth remains the only place in the Universe known to harbour life,^{[26][27]} and fossil evidence from the Earth informs most studies of abiogenesis. More than 99% of all species of life forms, amounting to over five billion species,^[28] that ever lived on Earth are estimated to be extinct.^{[29][30]} The age of the Earth is about 4.54 billion years,^{[31][32][33]} the earliest undisputed evidence of life on Earth dates from at least 3.5 billion years ago,^{[34][35][36]} and possibly as early as the Eoarchean Era (between 3.6 to 4.0 billion years ago), after geological crust started to solidify following the molten Hadean Eon. In May 2017, evidence of the earliest known life on land may have been found in 3.48-billion-year-old geysirite and other related mineral deposits (often found around hot springs and geysers) uncovered in the Pilbara Craton of Western Australia.^{[37][38]} However, there have been a number of discoveries that suggested the earliest appearance of life on Earth was even earlier in time. Currently, microfossils within hydrothermal vent precipitates dated from 3.77 to 4.28 billion years old found in Quebec, Canada may be the oldest record of life on Earth, suggesting "an almost instantaneous emergence of life" after ocean formation 4.4 billion years ago.^{[39][40][41][42][43]} According to biologist Stephen Blair Hedges, "If life arose relatively quickly on Earth ... then it could be common in the universe."^[44]

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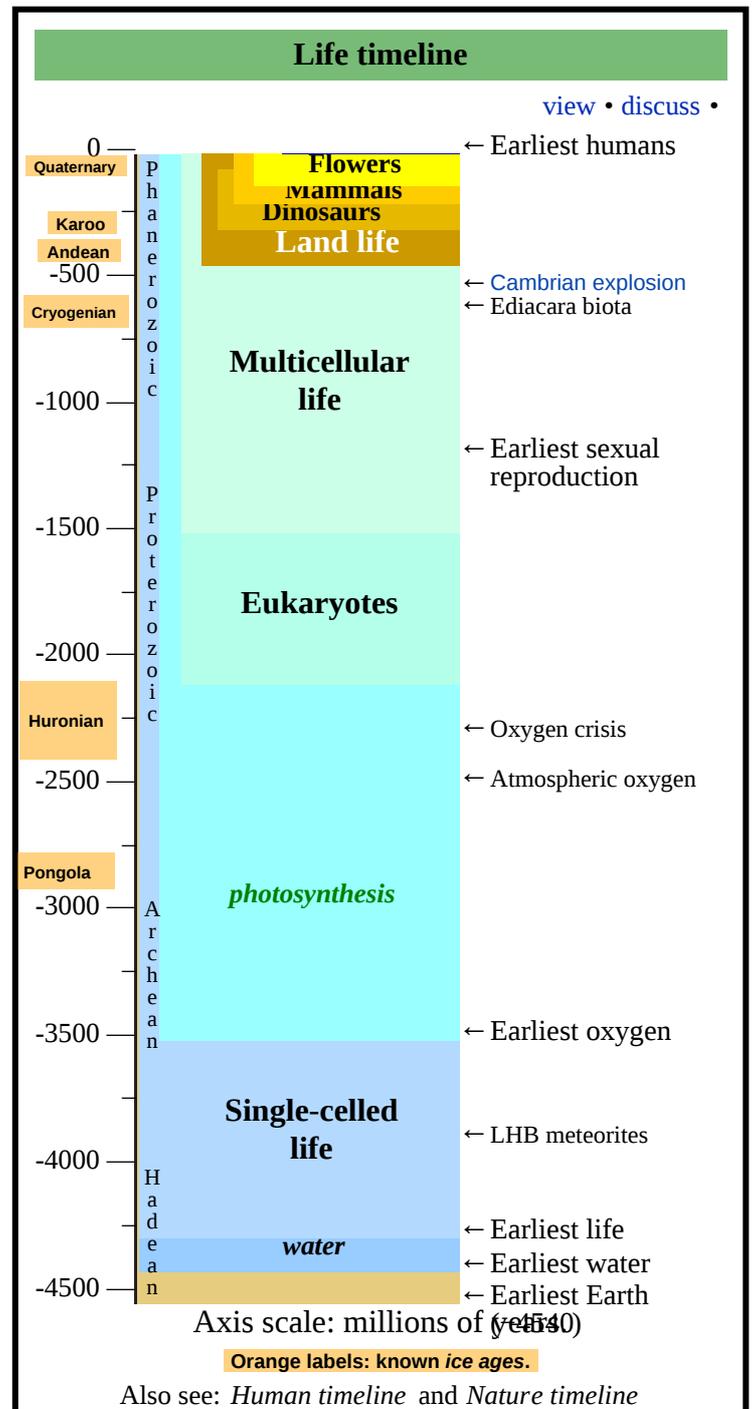
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Early geophysical conditions on Earth

The Hadean Earth is thought to have had a secondary atmosphere, formed through degassing of the rocks that accumulated from planetesimal impactors. At first, it was thought that the Earth's atmosphere consisted of hydrogen compounds—methane, ammonia and water vapour—and that life began under such reducing conditions, which are conducive to the formation of organic molecules. During its formation, the Earth lost a significant part of its initial mass, with a nucleus of the heavier rocky elements of the protoplanetary disk remaining.^[45] According to later models, suggested by study of ancient minerals, the atmosphere in the late Hadean period consisted largely of water vapour, nitrogen and carbon dioxide, with smaller amounts of carbon monoxide, hydrogen, and sulfur compounds.^[46] As Earth lacked the gravity to hold any molecular hydrogen, this component of the atmosphere would have been rapidly lost during the Hadean period, along with the bulk of the original inert gases. The solution of carbon dioxide in water is thought to have made the seas slightly acidic, giving it a pH of about 5.5. The atmosphere at the time has been characterized as a "gigantic, productive outdoor chemical laboratory."^[47] It may have been similar to the mixture of gases released today by volcanoes, which still support some abiotic chemistry.^[47]

Oceans may have appeared first in the Hadean Eon, as soon as two hundred million years (200 Ma) after the Earth was formed, in a hot 100 °C (212 °F) reducing environment, and the pH of about 5.8 rose rapidly towards neutral.^[48] This has been supported by the dating of 4.404 Ga-old zircon crystals from metamorphosed quartzite of Mount Narryer in the Western Australia Jack Hills of the Pilbara, which are evidence that oceans and continental crust existed within 150 Ma of Earth's formation.^[49] Despite the likely increased volcanism and existence of many smaller tectonic "platelets," it has been suggested that between 4.4 and 4.3 Ga (billion year), the Earth was a water world, with little if any continental crust, an extremely turbulent atmosphere and a hydrosphere subject to intense ultraviolet (UV) light, from a T Tauri stage Sun, cosmic radiation and continued bolide impacts.^[50]



The Hadean environment would have been highly hazardous to modern life. Frequent collisions with large objects, up to 500 kilometres (310 mi) in diameter, would have been sufficient to sterilize the planet and vaporize the ocean within a few months of impact, with hot steam mixed with rock vapour becoming high altitude clouds that would completely cover the planet. After a few months, the height of these clouds would have begun to decrease but the cloud base would still have been elevated for about the next thousand years. After that, it would have begun to rain at low altitude. For another two thousand years, rains would slowly have drawn down the height of the clouds, returning the oceans to their original depth only 3,000 years after the impact event.^[51]

Earliest biological evidence for life

The most commonly accepted location of the root of the tree of life is between a monophyletic domain Bacteria and a clade formed by Archaea and Eukaryota of what is referred to as the "traditional tree of life" based on several molecular studies starting with C. Woese.^[52] A very small minority of studies have concluded differently, namely that the root is in the Domain Bacteria, either in the phylum Firmicutes^[53] or that the phylum Chloroflexi is basal to a clade with Archaea+Eukaryotes and the rest of Bacteria as proposed by Thomas Cavalier-Smith.^[54] More recently Peter Ward has established an alternative view which is rooted in abiotic RNA synthesis which becomes enclosed within a capsule and then creates RNA ribosome replicates. It is proposed that this then bifurcates between Dominion Ribosa (hypothetical Domain Ribosa or RNA life), and after the loss of ribozymes RNA viruses as Domain Viorea, and Dominion Terroa, which after creating a large cell within a lipid wall, creating DNA the 20 based amino acids and the triplet code, is established as the last universal common ancestor or LUCA, of earlier phylogenetic trees.^[55]

The earliest life on Earth existed more than 3.5 billion years ago,^{[34][35][36]} during the Eoarchean Era when sufficient crust had solidified following the molten Hadean Eon. The earliest physical evidence so far found consists of microfossils in the Nuvvuagittuq Greenstone Belt of Northern Quebec, in "banded iron formation" rocks at least 3.77 billion and possibly 4.28 billion years old.^{[39][56]} This finding suggested that there was almost instant development of life after oceans were formed. The structure of the microbes was noted to be similar to bacteria found near hydrothermal vents in the modern era, and provided support for the hypothesis that abiogenesis began near hydrothermal vents.^{[42][39]}

Also noteworthy is biogenic graphite in 3.7 billion-year-old metasedimentary rocks from southwestern Greenland^[57] and microbial mat fossils found in 3.48 billion-year-old sandstone from Western Australia.^{[58][59]} Evidence of early life in rocks from Akilia Island, near the Isua supracrustal belt in southwestern Greenland, dating to 3.7 billion years ago have shown biogenic carbon isotopes.^{[60][61]} At Strelley Pool, in the Pilbara region of Western Australia, compelling evidence of early life was found in pyrite-bearing sandstone in a fossilized beach, that showed rounded tubular cells that oxidized sulfur by photosynthesis in the absence of oxygen.^{[62][63][64]} Further research on zircons from Western Australia in 2015 suggested evidence that life likely existed on Earth at least 4.1 billion years ago.^{[44][65][66]}

Traditionally it was thought that during the period between 3.8 and 4.1 Ga, changes in the orbits of the giant planets may have caused a heavy bombardment by asteroids and comets^[67] that pockmarked the Moon and the other inner planets (Mercury, Mars, and presumably Earth and Venus). This would likely have repeatedly sterilized the planet, had life appeared before that time.^[47] Geologically, the Hadean Earth would have been far more active than at any other time in its history. Studies of meteorites suggests that radioactive isotopes such as aluminium-26 with a half-life of 7.17×10^5 years, and potassium-40 with a half-life of 1.250×10^9 years, isotopes mainly produced in supernovae, were much more common.^[68] Internal heating as a result of gravitational sorting between the core and the mantle would have caused a great deal of mantle convection, with the probable result of many more smaller and more active tectonic plates than now exist.

The time periods between such devastating environmental events give time windows for the possible origin of life in the early environments. If the deep marine hydrothermal setting was the site for the origin of life, then abiogenesis could have happened as early as 4.0 to 4.2 Ga. If the site was at the surface of the Earth, abiogenesis could only have occurred between 3.7 and 4.0 Ga.^[69]

In 2016, a set of 355 genes likely present in the Last Universal Common Ancestor (LUCA) of all organisms living on Earth was identified.^[70] A total of 6.1 million prokaryotic protein coding genes from various phylogenetic trees were sequenced, identifying 355 protein clusters from amongst 286,514 protein clusters that were probably common to LUCA. The results "depict LUCA as anaerobic, CO₂-fixing, H₂-dependent with a Wood–Ljungdahl pathway, N₂-fixing and thermophilic. LUCA's biochemistry was replete with FeS clusters and radical reaction mechanisms. Its cofactors reveal dependence upon transition metals, flavins, S-adenosyl methionine, coenzyme A, ferredoxin, molybdopterin, corrins and selenium. Its genetic code required nucleoside modifications and S-adenosylmethionine-dependent methylations." The results depict methanogenic clostridia as a basal clade in the 355 phylogenies examined, and suggest that LUCA inhabited an anaerobic hydrothermal vent setting in a geochemically active environment rich in H₂, CO₂ and iron.^[71] M.D. Brazier has shown that the tiny fossils discovered came from a hot poisonous world of the toxic gases methane, ammonia, carbon dioxide and hydrogen sulphide.^[72] An analysis of the conventional three fold tree of life shows thermophilic and hyperthermophilic bacteria and archaea are closest to the root, suggesting that life may have evolved in a hot environment.^[73]

Conceptual history

Spontaneous generation

Belief in spontaneous generation of certain forms of life from non-living matter goes back to Aristotle and ancient Greek philosophy and continued to have support in Western scholarship until the 19th century.^[74] This belief was paired with a belief in heterogenesis, i.e., that one form of life derived from a different form (e.g., bees from flowers).^[75] Classical notions of spontaneous generation held that certain complex, living organisms are generated by decaying organic substances. According to Aristotle, it was a readily observable truth that aphids arise from the dew that falls on plants, flies from putrid matter, mice from dirty hay, crocodiles from rotting logs at the bottom of bodies of water, and so on.^[76] In the 17th century, people began to question such assumptions. In 1646, Sir Thomas Browne published his *Pseudodoxia Epidemica* (subtitled *Enquiries into Very many Received Tenets, and commonly Presumed Truths*), which was an attack on false beliefs and "vulgar errors." His contemporary, Alexander Ross, erroneously refuted him, stating: "To question this [Ed.: i.e., spontaneous generation], is to question Reason, Sense, and Experience: If he doubts of this, let him go to *Ægypt*, and there he will finde the fields swarming with mice begot of the mud of *Nylus*, to the great calamity of the Inhabitants."^{[77][78]}

In 1665, Robert Hooke published the first drawings of a microorganism. Hooke was followed in 1676 by Antonie van Leeuwenhoek, who drew and described microorganisms that are now thought to have been protozoa and bacteria.^[79] Many felt the existence of microorganisms was evidence in support of spontaneous generation, since microorganisms seemed too simplistic for sexual reproduction, and asexual reproduction through cell division had not yet been observed. Van Leeuwenhoek took issue with the ideas common at the time that fleas and lice could spontaneously result from putrefaction, and that frogs could likewise arise from slime. Using a broad range of experiments ranging from sealed and open meat incubation and the close study of insect reproduction he became, by the 1680s, convinced that spontaneous generation was incorrect.^[80]

The first experimental evidence against spontaneous generation came in 1668 when Francesco Redi showed that no maggots appeared in meat when flies were prevented from laying eggs. It was gradually shown that, at least in the case of all the higher and readily visible organisms, the previous sentiment regarding spontaneous

generation was false. The alternative seemed to be biogenesis: that every living thing came from a pre-existing living thing (*omne vivum ex ovo*, Latin for "every living thing from an egg").

In 1768, Lazzaro Spallanzani demonstrated that microbes were present in the air, and could be killed by boiling. In 1861, Louis Pasteur performed a series of experiments that demonstrated that organisms such as bacteria and fungi do not spontaneously appear in sterile, nutrient-rich media, but could only appear by invasion from without.

The belief that self-ordering by spontaneous generation was impossible begged for an alternative. By the middle of the 19th century, the theory of biogenesis had accumulated so much evidential support, due to the work of Pasteur and others, that the alternative theory of spontaneous generation had been effectively disproven. John Desmond Bernal, a pioneer in X-ray crystallography, suggested that earlier theories such as spontaneous generation were based upon an explanation that life was continuously created as a result of chance events.^[81]

Etymology

The term *biogenesis* is usually credited to either Henry Charlton Bastian or to Thomas Henry Huxley.^[82] Bastian used the term around 1869 in an unpublished exchange with John Tyndall to mean "life-origination or commencement". In 1870, Huxley, as new president of the British Association for the Advancement of Science, delivered an address entitled *Biogenesis and Abiogenesis*.^[83] In it he introduced the term *biogenesis* (with an opposite meaning to Bastian's) as well as *abiogenesis*:

And thus the hypothesis that living matter always arises by the agency of pre-existing living matter, took definite shape; and had, henceforward, a right to be considered and a claim to be refuted, in each particular case, before the production of living matter in any other way could be admitted by careful reasoners. It will be necessary for me to refer to this hypothesis so frequently, that, to save circumlocution, I shall call it the hypothesis of *Biogenesis*; and I shall term the contrary doctrine—that living matter may be produced by not living matter—the hypothesis of *Abiogenesis*.^[83]

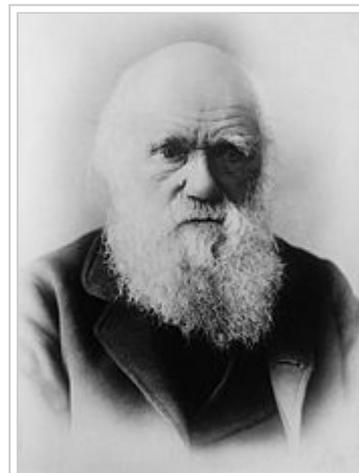
Subsequently, in the preface to Bastian's 1871 book, *The Modes of Origin of Lowest Organisms*,^[84] Bastian referred to the possible confusion with Huxley's usage and explicitly renounced his own meaning:

A word of explanation seems necessary with regard to the introduction of the new term *Archebiosis*. I had originally, in unpublished writings, adopted the word *Biogenesis* to express the same meaning—viz., life-origination or commencement. But in the mean time the word *Biogenesis* has been made use of, quite independently, by a distinguished biologist [Huxley], who wished to make it bear a totally different meaning. He also introduced the word *Abiogenesis*. I have been informed, however, on the best authority, that neither of these words can—with any regard to the language from which they are derived—be supposed to bear the meanings which have of late been publicly assigned to them. Wishing to avoid all needless confusion, I therefore renounced the use of the word *Biogenesis*, and being, for the reason just given, unable to adopt the other term, I was compelled to introduce a new word, in order to designate the process by which living matter is supposed to come into being, independently of pre-existing living matter.^[85]

Louis Pasteur and Charles Darwin

Louis Pasteur remarked, about a finding of his in 1864 which he considered definitive, "Never will the doctrine of spontaneous generation recover from the mortal blow struck by this simple experiment."^{[86][87]} One alternative was that life's origins on Earth had come from somewhere else in the Universe. Periodically resurrected (see Panspermia, above) Bernal said that this approach "is equivalent in the last resort to asserting the operation of metaphysical, spiritual entities... it turns on the argument of creation by design by a creator or demiurge."^[88] Such a theory, Bernal said, was unscientific. A theory popular around the same time was that life was the result of an inner "life force", which in the late 19th century was championed by Henri Bergson.

The idea of evolution by natural selection proposed by Charles Darwin put an end to these metaphysical theologies. In a letter to Joseph Dalton Hooker on 1 February 1871,^[89] Darwin discussed the suggestion that the original spark of life may have begun in a "warm little pond, with all sorts of ammonia and phosphoric salts, light, heat, electricity, &c., present, that a proteine compound was chemically formed ready to undergo still more complex changes." He went on to explain that "at the present day such matter would be instantly devoured or absorbed, which would not have been the case before living creatures were formed." He had written to Hooker in 1863 stating that, "It is mere rubbish, thinking at present of the origin of life; one might as well think of the origin of matter." In *On the Origin of Species*, he had referred to life having been "created", by which he "really meant 'appeared' by some wholly unknown process", but had soon regretted using the Old Testament term "creation".^[90]



Charles Darwin in 1879

"Primordial soup" hypothesis

No new notable research or theory on the subject appeared until 1924, when Alexander Oparin reasoned that atmospheric oxygen prevents the synthesis of certain organic compounds that are necessary building blocks for the evolution of life. In his book *The Origin of Life*,^{[91][92]} Oparin proposed that the "spontaneous generation of life" that had been attacked by Louis Pasteur did in fact occur once, but was now impossible because the conditions found on the early Earth had changed, and preexisting organisms would immediately consume any spontaneously generated organism. Oparin argued that a "primeval soup" of organic molecules could be created in an oxygenless atmosphere through the action of sunlight. These would combine in ever more complex ways until they formed coacervate droplets. These droplets would "grow" by fusion with other droplets, and "reproduce" through fission into daughter droplets, and so have a primitive metabolism in which factors that promote "cell integrity" survive, and those that do not become extinct. Many modern theories of the origin of life still take Oparin's ideas as a starting point.



Alexander Oparin (right) in his laboratory, 1938

Robert Shapiro has summarized the "primordial soup" theory of Oparin and J. B. S. Haldane in its "mature form" as follows:^[93]

1. The early Earth had a chemically reducing atmosphere.
2. This atmosphere, exposed to energy in various forms, produced simple organic compounds ("monomers").
3. These compounds accumulated in a "soup" that may have concentrated at various locations (shorelines, oceanic vents etc.).
4. By further transformation, more complex organic polymers – and ultimately life – developed in the soup.

About this time, Haldane suggested that the Earth's prebiotic oceans (quite different from their modern counterparts) would have formed a "hot dilute soup" in which organic compounds could have formed. Bernal called this idea *biopoiesis* or *biopoesis*, the process of living matter evolving from self-replicating but non-living molecules,^{[81][94]} and proposed that biopoiesis passes through a number of intermediate stages.

One of the most important pieces of experimental support for the "soup" theory came in 1952. Stanley L. Miller and Harold C. Urey performed an experiment that demonstrated how organic molecules could have spontaneously formed from inorganic precursors under conditions like those posited by the Oparin-Haldane hypothesis. The now-famous Miller–Urey experiment used a highly reducing mixture of gases – methane, ammonia, and hydrogen, as well as water vapour – to form basic organic monomers such as amino acids.^[95] The mixture of gases was cycled through an apparatus that delivered electrical sparks to the mixture. After one week, it was found that about 10% to 15% of the carbon in the system was then in the form of a racemic

mixture of organic compounds, including amino acids, which are the building blocks of proteins. This provided direct experimental support for the second point of the "soup" theory, and it is around the remaining two points of the theory that much of the debate now centres.

Bernal showed that based upon this and subsequent work there is no difficulty in principle in forming most of the molecules we recognize as the basic molecules of life from their inorganic precursors. The underlying hypothesis held by Oparin, Haldane, Bernal, Miller and Urey, for instance, was that multiple conditions on the primeval Earth favoured chemical reactions that synthesized the same set of complex organic compounds from such simple precursors. A 2011 reanalysis of the saved vials containing the original extracts that resulted from the Miller and Urey experiments, using current and more advanced analytical equipment and technology, has uncovered more biochemicals than originally discovered in the 1950s. One of the more important findings was 23 amino acids, far more than the five originally found.^[96] However, Bernal said that "it is not enough to explain the formation of such molecules, what is necessary, is a physical-chemical explanation of the origins of these molecules that suggests the presence of suitable sources and sinks for free energy."^[97]

Proteinoid microspheres

In trying to uncover the intermediate stages of abiogenesis mentioned by Bernal, Sidney W. Fox in the 1950s and 1960s studied the spontaneous formation of peptide structures under conditions that might plausibly have existed early in Earth's history. He demonstrated that amino acids could spontaneously form small chains called peptides. In one of his experiments, he allowed amino acids to dry out as if puddled in a warm, dry spot in prebiotic conditions. He found that, as they dried, the amino acids formed long, often cross-linked, thread-like, submicroscopic polypeptide molecules now named "proteinoid microspheres."^[98]

In another experiment using a similar method to set suitable conditions for life to form, Fox collected volcanic material from a cinder cone in Hawaii. He discovered that the temperature was over 100 °C (212 °F) just 4 inches (100 mm) beneath the surface of the cinder cone, and suggested that this might have been the environment in which life was created—molecules could have formed and then been washed through the loose volcanic ash and into the sea. He placed lumps of lava over amino acids derived from methane, ammonia and water, sterilized all materials, and baked the lava over the amino acids for a few hours in a glass oven. A brown, sticky substance formed over the surface and when the lava was drenched in sterilized water a thick, brown liquid leached out. It turned out that the amino acids had combined to form proteinoids, and the proteinoids had combined to form small globules that Fox called "microspheres." His proteinoids were not cells, although they formed clumps and chains reminiscent of cyanobacteria, but they contained no functional nucleic acids or any encoded information. Based upon such experiments, Colin S. Pittendrigh stated in December 1967 that "laboratories will be creating a living cell within ten years," a remark that reflected the typical contemporary levels of innocence of the complexity of cell structures.^[99]

Current models

There is no single, generally accepted model for the origin of life. Scientists have proposed several plausible theories, which share some common elements. While differing in the details, these theories are based on the framework laid out by Alexander Oparin (in 1924) and by J. B. S. Haldane (in 1925), who postulated the molecular or chemical evolution theory of life.^[100] According to them, the first molecules constituting the earliest cells "were synthesized under natural conditions by a slow process of molecular evolution, and these molecules then organized into the first molecular system with properties with biological order".^[100] Oparin and Haldane suggested that the atmosphere of the early Earth may have been chemically reducing in nature, composed primarily of methane (CH₄), ammonia (NH₃), water (H₂O), hydrogen sulfide (H₂S), carbon dioxide (CO₂) or carbon monoxide (CO), and phosphate (PO₄³⁻), with molecular oxygen (O₂) and ozone (O₃) either rare or absent. According to later models, the atmosphere in the late Hadean period consisted largely of nitrogen (N₂) and carbon dioxide, with smaller amounts of carbon monoxide, hydrogen (H₂), and sulfur compounds;^[101] while it did lack molecular oxygen and ozone,^[102] it was not as chemically reducing as

Oparin and Haldane supposed. In the atmosphere proposed by Oparin and Haldane, electrical activity can produce certain basic small molecules (monomers) of life, such as amino acids. The Miller–Urey experiment reported in 1953 demonstrated this.

Bernal coined the term *biopoiesis* in 1949 to refer to the origin of life.^[103] In 1967, he suggested that it occurred in three "stages":

1. the origin of biological monomers
2. the origin of biological polymers
3. the evolution from molecules to cells

Bernal suggested that evolution commenced between stages 1 and 2. Bernal regarded the third stage – discovering methods by which biological reactions were incorporated behind a cell's boundary – as the most difficult. Modern work on the way that cell membranes self-assemble, and the work on micropores in various substrates may be a halfway house towards the development of independent free-living cells.^{[104][105][106]}

The chemical processes that took place on the early Earth are called *chemical evolution*. Both Manfred Eigen and Sol Spiegelman demonstrated that evolution, including replication, variation, and natural selection, can occur in populations of molecules as well as in organisms.^[47] Spiegelman took advantage of natural selection to synthesize the Spiegelman Monster, which had a genome with just 218 nucleotide bases, having deconstructively evolved from a 4500-base bacterial RNA. Eigen built on Spiegelman's work and produced a similar system further degraded to just 48 or 54 nucleotides – the minimum required for the binding of the replication enzyme.^[107]

Following on from chemical evolution came the initiation of biological evolution, which led to the first cells.^[47] No one has yet synthesized a "protocell" using basic components with the necessary properties of life (the so-called "bottom-up-approach"). Without such a proof-of-principle, explanations have tended to focus on chemosynthesis.^[108] However, some researchers work in this field, notably Steen Rasmussen and Jack W. Szostak. Others have argued that a "top-down approach" is more feasible. One such approach, successfully attempted by Craig Venter and others at J. Craig Venter Institute, involves engineering existing prokaryotic cells with progressively fewer genes, attempting to discern at which point the most minimal requirements for life are reached.^{[109][110][111]}

Chemical origin of organic molecules

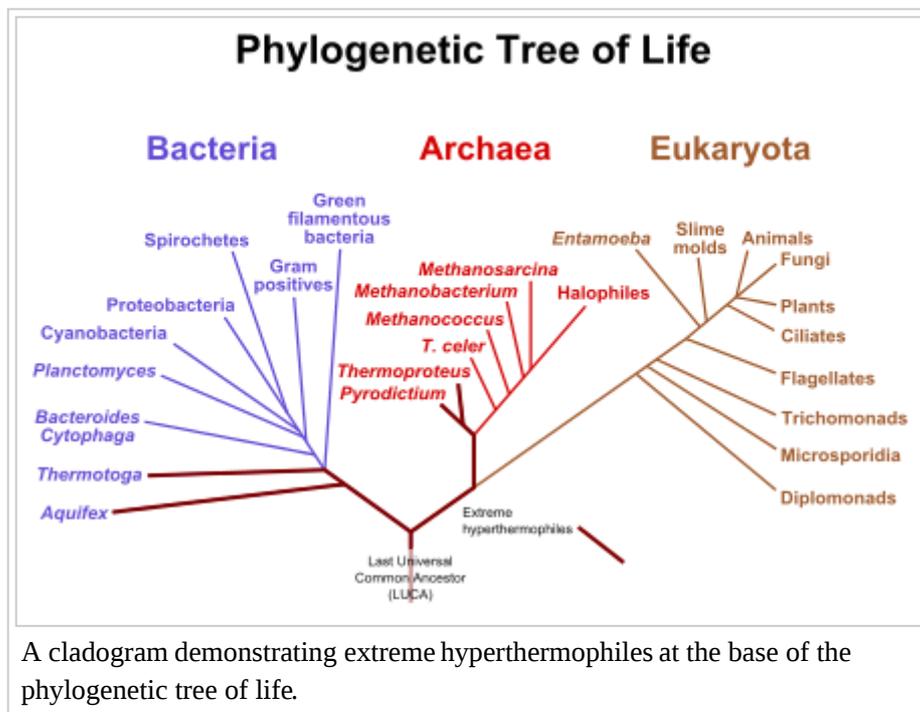
The elements, except for hydrogen and helium, ultimately derive from stellar nucleosynthesis. On 12 October 2016, astronomers reported that the very basic chemical ingredients of life—the carbon-hydrogen molecule (CH, or methylidyne radical), the carbon-hydrogen positive ion (CH⁺) and the carbon ion (C⁺)—are the result, in large part, of ultraviolet light from stars, rather than in other ways, such as the result of turbulent events related to supernovae and young stars, as thought earlier.^[112] Complex molecules, including organic molecules, form naturally both in space and on planets.^[19] There are two possible sources of organic molecules on the early Earth:

1. Terrestrial origins – organic molecule synthesis driven by impact shocks or by other energy sources (such as UV light, redox coupling, or electrical discharges) (e.g., Miller's experiments)
2. Extraterrestrial origins – formation of organic molecules in interstellar dust clouds, which rain down on planets.^{[113][114]} (See pseudo-panspermia)

Based on recent computer model studies, the complex organic molecules necessary for life may have formed in the protoplanetary disk of dust grains surrounding the Sun before the formation of the Earth.^{[115][116]} According to the computer studies, this same process may also occur around other stars that acquire planets. (Also see Extraterrestrial organic molecules).

Estimates of the production of organics from these sources suggest that the Late Heavy Bombardment before 3.5 Ga within the early atmosphere made available quantities of organics comparable to those produced by terrestrial sources.^{[117][118]}

It has been estimated that the Late Heavy Bombardment may also have effectively sterilized the Earth's surface to a depth of tens of metres. If life evolved deeper than this, it would have also been shielded from the early high levels of ultraviolet radiation from the T Tauri stage of the Sun's evolution. Simulations of geothermally heated oceanic crust yield far more organics than those found in the Miller-Urey experiments (see below). In the deep hydrothermal vents, Everett Shock has found "there is an enormous thermodynamic drive to form organic compounds, as seawater and hydrothermal fluids, which are far from equilibrium, mix and move towards a more stable state."^[119] Shock has found that the available energy is maximized at around 100 – 150 degrees Celsius, precisely the temperatures at which the hyperthermophilic bacteria and thermoacidophilic archaea have been found, at the base of the phylogenetic tree of life closest to the Last Universal Common Ancestor (LUCA).^[120]



Chemical synthesis

While features of self-organization and self-replication are often considered the hallmark of living systems, there are many instances of abiotic molecules exhibiting such characteristics under proper conditions. Stan Palasek suggested based on a theoretical model that self-assembly of ribonucleic acid (RNA) molecules can occur spontaneously due to physical factors in hydrothermal vents.^[121] Virus self-assembly within host cells has implications for the study of the origin of life,^[122] as it lends further credence to the hypothesis that life could have started as self-assembling organic molecules.^{[123][124]}

Multiple sources of energy were available for chemical reactions on the early Earth. For example, heat (such as from geothermal processes) is a standard energy source for chemistry. Other examples include sunlight and electrical discharges (lightning), among others.^[47] Unfavourable reactions can also be driven by highly favourable ones, as in the case of iron-sulfur chemistry. For example, this was probably important for carbon fixation (the conversion of carbon from its inorganic form to an organic one).^[note 1] Carbon fixation via iron-sulfur chemistry is highly favourable, and occurs at neutral pH and 100 °C (212 °F). Iron-sulfur surfaces, which are abundant near hydrothermal vents, are also capable of producing small amounts of amino acids and other biological metabolites.^[47]

Formamide produces all four ribonucleotides and other biological molecules when warmed in the presence of various terrestrial minerals. Formamide is ubiquitous in the Universe, produced by the reaction of water and hydrogen cyanide (HCN). It has several advantages as a biotic precursor, including the ability to easily become concentrated through the evaporation of water.^{[125][126]} Although HCN is poisonous, it only affects aerobic organisms (eukaryotes and aerobic bacteria), which did not yet exist. It can play roles in other chemical processes as well, such as the synthesis of the amino acid glycine.^[47]

In 1961, it was shown that the nucleic acid purine base adenine can be formed by heating aqueous ammonium cyanide solutions.^[127] Other pathways for synthesizing bases from inorganic materials were also reported.^[128] Leslie E. Orgel and colleagues have shown that freezing temperatures are advantageous for the synthesis of purines, due to the concentrating effect for key precursors such as hydrogen cyanide.^[129] Research by Stanley L. Miller and colleagues suggested that while adenine and guanine require freezing conditions for synthesis, cytosine and uracil may require boiling temperatures.^[130] Research by the Miller group notes the formation of seven different amino acids and 11 types of nucleobases in ice when ammonia and cyanide were left in a freezer from 1972 to 1997.^{[131][132]} Other work demonstrated the formation of s-triazines (alternative nucleobases), pyrimidines (including cytosine and uracil), and adenine from urea solutions subjected to freeze-thaw cycles under a reductive atmosphere (with spark discharges as an energy source).^[133] The explanation given for the unusual speed of these reactions at such a low temperature is eutectic freezing. As an ice crystal forms, it stays pure: only molecules of water join the growing crystal, while impurities like salt or cyanide are excluded. These impurities become crowded in microscopic pockets of liquid within the ice, and this crowding causes the molecules to collide more often. Mechanistic exploration using quantum chemical methods provide a more detailed understanding of some of the chemical processes involved in chemical evolution, and a partial answer to the fundamental question of molecular biogenesis.^[134]

At the time of the Miller–Urey experiment, scientific consensus was that the early Earth had a reducing atmosphere with compounds relatively rich in hydrogen and poor in oxygen (e.g., CH₄ and NH₃ as opposed to CO₂ and nitrogen dioxide (NO₂)). However, current scientific consensus describes the primitive atmosphere as either weakly reducing or neutral^{[135][136]} (see also Oxygen Catastrophe). Such an atmosphere would diminish both the amount and variety of amino acids that could be produced, although studies that include iron and carbonate minerals (thought present in early oceans) in the experimental conditions have again produced a diverse array of amino acids.^[135] Other scientific research has focused on two other potential reducing environments: outer space and deep-sea thermal vents.^{[137][138][139]}

The spontaneous formation of complex polymers from abiotically generated monomers under the conditions posited by the "soup" theory is not at all a straightforward process. Besides the necessary basic organic monomers, compounds that would have prohibited the formation of polymers were also formed in high concentration during the Miller–Urey and Joan Oró experiments.^[140] The Miller–Urey experiment, for example, produces many substances that would react with the amino acids or terminate their coupling into peptide chains.^[141]

A research project completed in March 2015 by John D. Sutherland and others found that a network of reactions beginning with hydrogen cyanide and hydrogen sulfide, in streams of water irradiated by UV light, could produce the chemical components of proteins and lipids, as well as those of RNA,^{[142][143]} while not producing a wide range of other compounds.^[144] The researchers used the term "cyanosulfidic" to describe this network of reactions.^[143]

Autocatalysis

Autocatalysts are substances that catalyze the production of themselves and therefore are "molecular replicators." The simplest self-replicating chemical systems are autocatalytic, and typically contain three components: a product molecule and two precursor molecules. The product molecule joins together the precursor molecules, which in turn produce more product molecules from more precursor molecules. The product molecule catalyzes the reaction by providing a complementary template that binds to the precursors, thus bringing them together. Such systems have been demonstrated both in biological macromolecules and in small organic molecules.^{[145][146]} Systems that do not proceed by template mechanisms, such as the self-reproduction of micelles and vesicles, have also been observed.^[146]

It has been proposed that life initially arose as autocatalytic chemical networks.^[147] British ethologist Richard Dawkins wrote about autocatalysis as a potential explanation for the origin of life in his 2004 book *The Ancestor's Tale*.^[148] In his book, Dawkins cites experiments performed by Julius Rebek, Jr. and his colleagues in which they combined amino adenosine and pentafluorophenyl esters with the autocatalyst amino adenosine triacid ester (AATE). One product was a variant of AATE, which catalyzed the synthesis of themselves. This experiment demonstrated the possibility that autocatalysts could exhibit competition within a population of entities with heredity, which could be interpreted as a rudimentary form of natural selection.^{[149][150]}

In the early 1970s, Manfred Eigen and Peter Schuster examined the transient stages between the molecular chaos and a self-replicating hypercycle in a prebiotic soup.^[151] In a hypercycle, the information storing system (possibly RNA) produces an enzyme, which catalyzes the formation of another information system, in sequence until the product of the last aids in the formation of the first information system. Mathematically treated, hypercycles could create quasispecies, which through natural selection entered into a form of Darwinian evolution. A boost to hypercycle theory was the discovery of ribozymes capable of catalyzing their own chemical reactions. The hypercycle theory requires the existence of complex biochemicals, such as nucleotides, which do not form under the conditions proposed by the Miller–Urey experiment.

Geoffrey W. Hoffmann has shown that an early error-prone translation machinery can be stable against an error catastrophe of the type that had been envisaged as problematical for the origin of life, and was known as "Orgel's paradox".^{[152][153][154]}

Hoffmann has furthermore argued that a complex nucleation event as the origin of life involving both polypeptides and nucleic acid is compatible with the time and space available in the primitive oceans of Earth.^[155] Hoffmann suggests that volcanic ash may provide the many random shapes needed in the postulated complex nucleation event. This aspect of the theory can be tested experimentally.

Homochirality

Homochirality refers to the geometric property of some materials that are composed of chiral units. Chiral refers to nonsuperimposable 3D forms that are mirror images of one another, as are left and right hands. Living organisms use molecules that have the same chirality ("handedness"): with almost no exceptions,^[156] amino acids are left-handed while nucleotides and sugars are right-handed. Chiral molecules can be synthesized, but in the absence of a chiral source or a chiral catalyst, they are formed in a 50/50 mixture of both enantiomers (called a racemic mixture). Known mechanisms for the production of non-racemic mixtures from racemic starting materials include: asymmetric physical laws, such as the electroweak interaction; asymmetric environments, such as those caused by circularly polarized light, quartz crystals, or the Earth's rotation; and statistical fluctuations during racemic synthesis.^[157]

Once established, chirality would be selected for.^[158] A small bias (enantiomeric excess) in the population can be amplified into a large one by asymmetric autocatalysis, such as in the Soai reaction.^[159] In asymmetric autocatalysis, the catalyst is a chiral molecule, which means that a chiral molecule is catalyzing its own production. An initial enantiomeric excess, such as can be produced by polarized light, then allows the more abundant enantiomer to outcompete the other.^[160]

Clark has suggested that homochirality may have started in outer space, as the studies of the amino acids on the Murchison meteorite showed that L-alanine is more than twice as frequent as its D form, and L-glutamic acid was more than three times prevalent than its D counterpart. Various chiral crystal surfaces can also act as sites for possible concentration and assembly of chiral monomer units into macromolecules.^[161] Compounds found on meteorites suggest that the chirality of life derives from abiogenic synthesis, since amino acids from meteorites show a left-handed bias, whereas sugars show a predominantly right-handed bias, the same as found in living organisms.^[162]

Self-enclosurement, reproduction, duplication and the RNA world

Protocells

A protocell is a self-organized, self-ordered, spherical collection of lipids proposed as a stepping-stone to the origin of life.^[163] A central question in evolution is how simple protocells first arose and differed in reproductive contribution to the following generation driving the evolution of life. Although a functional protocell has not yet been achieved in a laboratory setting, there are scientists who think the goal is well within reach.^{[164][165][166]}

Self-assembled vesicles are essential components of primitive cells.^[163] The second law of thermodynamics requires that the Universe move in a direction in which entropy increases, yet life is distinguished by its great degree of organization. Therefore, a boundary is needed to separate life processes from non-living matter.^[167] Researchers Irene A. Chen and Jack W. Szostak amongst others, suggest that simple physicochemical properties of elementary protocells can give rise to essential cellular behaviours, including primitive forms of differential reproduction competition and energy storage. Such cooperative interactions between the membrane and its encapsulated contents could greatly simplify the transition from simple replicating molecules to true cells.^[165] Furthermore, competition for membrane molecules would favour stabilized membranes, suggesting a selective advantage for the evolution of cross-linked fatty acids and even the phospholipids of today.^[165] Such micro-encapsulation would allow for metabolism within the membrane, the exchange of small molecules but the prevention of passage of large substances across it.^[168] The main advantages of encapsulation include the increased solubility of the contained cargo within the capsule and the storage of energy in the form of an electrochemical gradient.

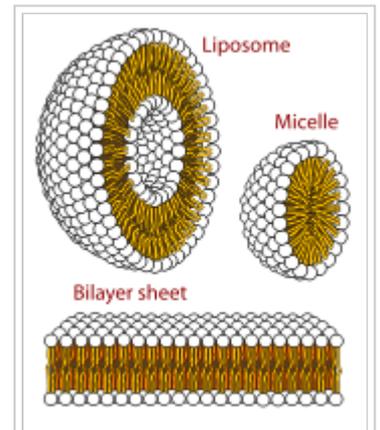
A 2012 study led by Armen Y. Mulkiđjanian of Germany's University of Osnabrück, suggests that inland pools of condensed and cooled geothermal vapour have the ideal characteristics for the origin of life.^[169] Scientists confirmed in 2002 that by adding a montmorillonite clay to a solution of fatty acid micelles (lipid spheres), the clay sped up the rate of vesicles formation 100-fold.^[166]

Another protocell model is the Jeewanu. First synthesized in 1963 from simple minerals and basic organics while exposed to sunlight, it is still reported to have some metabolic capabilities, the presence of semipermeable membrane, amino acids, phospholipids, carbohydrates and RNA-like molecules.^{[170][171]} However, the nature and properties of the Jeewanu remains to be clarified.

Electrostatic interactions induced by short, positively charged, hydrophobic peptides containing 7 amino acids in length or fewer, can attach RNA to a vesicle membrane, the basic cell membrane.^[172]

RNA world

The RNA world hypothesis describes an early Earth with self-replicating and catalytic RNA but no DNA or proteins.^[174] It is generally accepted that current life on Earth descends from an RNA world,^{[15][175]} although RNA-based life may not have been the first life to exist.^{[16][17]} This conclusion is drawn from many independent lines of evidence, such as the observations that RNA is central to the translation process and that small RNAs can catalyze all of the chemical groups and information transfers required for life.^{[17][176]} The structure of the ribosome has been called the "smoking gun," as it showed that the ribosome is a ribozyme, with a central core of RNA and no amino acid side chains within 18 angstroms of the active site where peptide bond



The three main structures phospholipids form spontaneously in solution: the liposome (a closed bilayer), the micelle and the bilayer.

formation is catalyzed.^[16] The concept of the RNA world was first proposed in 1962 by Alexander Rich,^[177] and the term was coined by Walter Gilbert in 1986.^{[17][178]}

Possible precursors for the evolution of protein synthesis include a mechanism to synthesize short peptide cofactors or form a mechanism for the duplication of RNA. It is likely that the ancestral ribosome was composed entirely of RNA, although some roles have since been taken over by proteins. Major remaining questions on this topic include identifying the selective force for the evolution of the ribosome and determining how the genetic code arose.^[179]

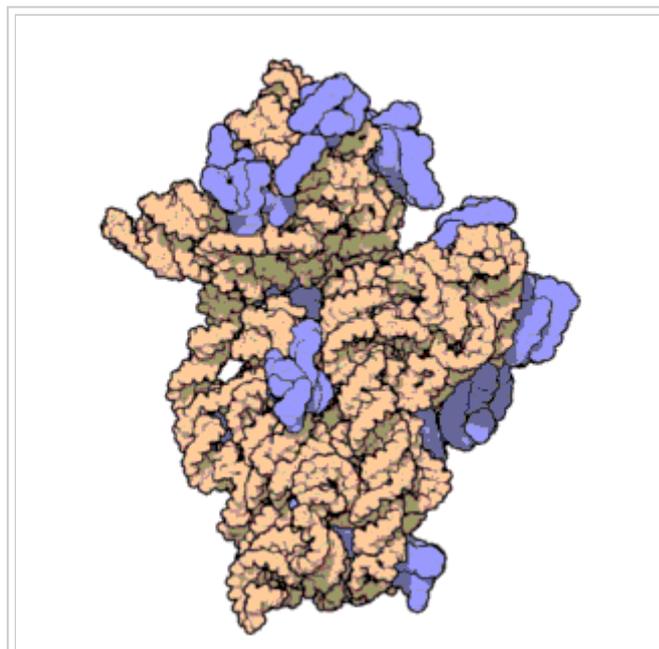
Eugene Koonin said, "Despite considerable experimental and theoretical effort, no compelling scenarios currently exist for the origin of replication and translation, the key processes that together comprise the core of biological systems and the apparent pre-requisite of biological evolution. The RNA World concept might offer the best chance for the resolution of this conundrum but so far cannot adequately account for the emergence of an efficient RNA replicase or the translation system. The MWO [Ed.: "many worlds in one"] version of the cosmological model of eternal inflation could suggest a way out of this conundrum because, in an infinite multiverse with a finite number of distinct macroscopic histories (each repeated an infinite number of times), emergence of even highly complex systems by chance is not just possible but inevitable."^[180]

Viral origins and the RNA World

Recent evidence for a "virus first" hypothesis, which may support theories of the RNA world have been suggested in new research.^[181] One of the difficulties for the study of viral origins and evolution is their high rate of mutation; this is particularly the case in RNA retroviruses like HIV.^[182] A 2015 study compared protein fold structures across different branches of the tree of life, where researchers can reconstruct the evolutionary histories of the folds and of the organisms whose genomes code for those folds. They argue that protein folds are better markers of ancient events as their three-dimensional structures can be maintained even as the sequences that code for those begin to change.^[181] Thus, the viral protein repertoire retain traces of ancient evolutionary history that can be recovered using advanced bioinformatics approaches. Those researchers think that "the prolonged pressure of genome and particle size reduction eventually reduced virocells into modern viruses (identified by the complete loss of cellular makeup), meanwhile other coexisting cellular lineages diversified into modern cells.^[183] The data suggest that viruses originated from ancient cells that co-existed with the ancestors of modern cells.^[181] These ancient cells likely contained segmented RNA genomes.^{[181][184]}

RNA synthesis and replication

The RNA world hypothesis has spurred scientists to determine if RNA molecules could have spontaneously formed able to catalyze their own replication.^{[185][186][187]} Evidence suggests that the chemical conditions, including the presence of boron, molybdenum and oxygen needed for the initial production of RNA molecules, may have been better on the planet Mars than on the planet Earth.^{[185][186]} If so, life-suitable molecules originating on Mars, may have later migrated to Earth via meteor ejections.^{[185][186]}



Molecular structure of the ribosome 30S subunit from *Thermus thermophilus*.^[173] Proteins are shown in blue and the single RNA chain in orange.

A number of hypotheses of formation of RNA have been put forward. As of 1994, there were difficulties in the explanation of the abiotic synthesis of the nucleotides cytosine and uracil.^[188] Subsequent research has shown possible routes of synthesis; for example, formamide produces all four ribonucleotides and other biological molecules when warmed in the presence of various terrestrial minerals.^{[125][126]} Early cell membranes could have formed spontaneously from proteinoids, which are protein-like molecules produced when amino acid solutions are heated while in the correct concentration of aqueous solution. These are seen to form microspheres which are observed to behave similarly to membrane-enclosed compartments. Other possible means of producing more complicated organic molecules include chemical reactions that take place on clay substrates or on the surface of the mineral pyrite.

Factors supportive of an important role for RNA in early life include its ability to act both to store information and to catalyze chemical reactions (as a ribozyme); its many important roles as an intermediate in the expression of and maintenance of the genetic information (in the form of DNA) in modern organisms; and the ease of chemical synthesis of at least the components of the RNA molecule under the conditions that approximated the early Earth. Relatively short RNA molecules have been artificially produced in labs, which are capable of replication.^[189] Such replicase RNA, which functions as both code and catalyst provides its own template upon which copying can occur. Jack W. Szostak has shown that certain catalytic RNAs can join smaller RNA sequences together, creating the potential for self-replication. If these conditions were present, Darwinian natural selection would favour the proliferation of such autocatalytic sets, to which further functionalities could be added.^[190] Such autocatalytic systems of RNA capable of self-sustained replication have been identified.^[191] The RNA replication systems, which include two ribozymes that catalyze each other's synthesis, showed a doubling time of the product of about one hour, and were subject to natural selection under the conditions that existed in the experiment.^[192] In evolutionary competition experiments, this led to the emergence of new systems which replicated more efficiently.^[16] This was the first demonstration of evolutionary adaptation occurring in a molecular genetic system.^[192]

Depending on the specific definition used, life can be considered to have emerged when RNA chains began to express the basic conditions necessary for natural selection to operate as conceived by Darwin: heritability, variation of type, and differential reproductive output. The fitness of an RNA replicator (its per capita rate of increase) would likely be a function of its adaptive capacities that are intrinsic (in the sense that they were determined by the nucleotide sequence) and the availability of its resources.^{[193][194]} The three primary adaptive capacities may have been (1) the capacity to replicate with moderate fidelity, giving rise to both heritability while allowing variation of type, (2) the capacity to avoid decay, and (3) the capacity to acquire and process resources.^{[193][194]} These capacities would have been determined initially by the folded configurations of the RNA replicators that, in turn, would be encoded in their individual nucleotide sequences. Relative reproductive success, competition, between different replicators would have depended on the relative values of their adaptive capacities.

Pre-RNA world

It is possible that a different type of nucleic acid, such as PNA, TNA or GNA, was the first to emerge as a self-reproducing molecule, only later replaced by RNA.^{[195][196]} Larralde *et al.*, say that "the generally accepted prebiotic synthesis of ribose, the formose reaction, yields numerous sugars without any selectivity."^[197] and they conclude that their "results suggest that the backbone of the first genetic material could not have contained ribose or other sugars because of their instability." The ester linkage of ribose and phosphoric acid in RNA is known to be prone to hydrolysis.^[198]

Pyrimidine ribonucleosides and their respective nucleotides have been prebiotically synthesized by a sequence of reactions which by-pass the free sugars, and are assembled in a stepwise fashion by using nitrogenous or oxygenous chemistries. Sutherland has demonstrated high yielding routes to cytidine and uridine ribonucleotides built from small 2 and 3 carbon fragments such as glycolaldehyde, glyceraldehyde or glyceraldehyde-3-phosphate, cyanamide and cyanoacetylene. One of the steps in this sequence allows the

isolation of enantiopure ribose aminooxazoline if the enantiomeric excess of glyceraldehyde is 60% or greater.^[199] This can be viewed as a prebiotic purification step, where the said compound spontaneously crystallized out from a mixture of the other pentose aminooxazolines. Ribose aminooxazoline can then react with cyanoacetylene in a mild and highly efficient manner to give the alpha cytidine ribonucleotide. Photoanomerization with UV light allows for inversion about the 1' anomeric centre to give the correct beta stereochemistry.^[200] In 2009 they showed that the same simple building blocks allow access, via phosphate controlled nucleobase elaboration, to 2',3'-cyclic pyrimidine nucleotides directly, which are known to be able to polymerize into RNA. This paper also highlights the possibility for the photo-sanitization of the pyrimidine-2',3'-cyclic phosphates.^[201]

Origin of biological metabolism

Metabolism-like reactions could have occurred naturally in early oceans, before the first organisms evolved.^{[18][202]} Metabolism may predate the origin of life and life may have evolved from the chemical conditions that prevailed in the world's earliest oceans. Reconstructions in laboratories show that some of these reactions can produce RNA, and some others resemble two essential reaction cascades of metabolism: glycolysis and the pentose phosphate pathway, that provide essential precursors for nucleic acids, amino acids and lipids.^[202] A study at the University of Düsseldorf created phylogenetic trees based upon 6 million genes from bacteria and archaea, and identified 355 protein families that were probably present in LUCA. They were based upon an anaerobic metabolism fixing carbon dioxide and nitrogen. It suggests that LUCA evolved in an environment rich in hydrogen, carbon dioxide and iron.^[203] Following are some observed discoveries and related hypotheses.

Iron–sulfur world

In the 1980s, Günter Wächtershäuser, encouraged and supported by Karl R. Popper,^{[204][205][206]} postulated in his iron–sulfur world, a theory of the evolution of pre-biotic chemical pathways as the starting point in the evolution of life. It presents a consistent system of tracing today's biochemistry back to ancestral reactions that provide alternative pathways to the synthesis of organic building blocks from simple gaseous compounds.

In contrast to the classical Miller experiments, which depend on external sources of energy (such as simulated lightning or ultraviolet irradiation), "Wächtershäuser systems" come with a built-in source of energy, sulfides of iron (iron pyrite) and other minerals. The energy released from redox reactions of these metal sulfides is available for the synthesis of organic molecules. It is therefore hypothesized that such systems may be able to evolve into autocatalytic sets of self-replicating, metabolically active entities that predate the life forms known today.^{[18][202]} Experiments with such sulfides in an aqueous environment at 100 °C produced a relatively small yield of dipeptides (0.4% to 12.4%) and a smaller yield of tripeptides (0.10%) although under the same conditions, dipeptides were quickly broken down.^[207]

Several models reject the idea of the self-replication of a "naked-gene" but postulate the emergence of a primitive metabolism which could provide a safe environment for the later emergence of RNA replication. The centrality of the Krebs cycle (citric acid cycle) to energy production in aerobic organisms, and in drawing in carbon dioxide and hydrogen ions in biosynthesis of complex organic chemicals, suggests that it was one of the first parts of the metabolism to evolve.^[208] Somewhat in agreement with these notions, geochemist Michael Russell has proposed that "the purpose of life is to hydrogenate carbon dioxide" (as part of a "metabolism-first," rather than a "genetics-first," scenario).^{[209][210]} Physicist Jeremy England of MIT has proposed that thermodynamically, life was bound to eventually arrive, as based on established physics, he mathematically indicates "...that when a group of atoms is driven by an external source of energy (like the sun or chemical fuel) and surrounded by a heat bath (like the ocean or atmosphere), it will often gradually restructure itself in order to dissipate increasingly more energy. This could mean that under certain conditions, matter inexorably acquires the key physical attribute associated with life."^{[211][212]}

One of the earliest incarnations of this idea was put forward in 1924 with Oparin's notion of primitive self-replicating vesicles which predated the discovery of the structure of DNA. Variants in the 1980s and 1990s include Wächtershäuser's iron-sulfur world theory and models introduced by Christian de Duve based on the chemistry of thioesters. More abstract and theoretical arguments for the plausibility of the emergence of metabolism without the presence of genes include a mathematical model introduced by Freeman Dyson in the early 1980s and Stuart Kauffman's notion of collectively autocatalytic sets, discussed later in that decade.

Orgel summarized his analysis of the proposal by stating, "There is at present no reason to expect that multistep cycles such as the reductive citric acid cycle will self-organize on the surface of FeS/FeS₂ or some other mineral."^[213] It is possible that another type of metabolic pathway was used at the beginning of life. For example, instead of the reductive citric acid cycle, the "open" acetyl-CoA pathway (another one of the five recognized ways of carbon dioxide fixation in nature today) would be compatible with the idea of self-organization on a metal sulfide surface. The key enzyme of this pathway, carbon monoxide dehydrogenase/acetyl-CoA synthase harbours mixed nickel-iron-sulfur clusters in its reaction centres and catalyzes the formation of acetyl-CoA (which may be regarded as a modern form of acetyl-thiol) in a single step. There are increasing concerns, however, that prebiotic thiolated (i.e. Thioacetic acid) and Thioester compounds are thermodynamically and kinetically unfavourable to accumulate in presumed prebiotic conditions (i.e. Hydrothermal vents).^[214]

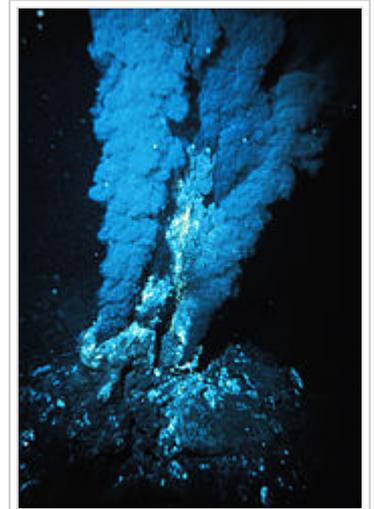
Zn-world hypothesis

The Zn-world (zinc world) theory of Armen Y. Mulkidjanian^[215] is an extension of Wächtershäuser's pyrite hypothesis. Wächtershäuser based his theory of the initial chemical processes leading to informational molecules (i.e., RNA, peptides) on a regular mesh of electric charges at the surface of pyrite that may have made the primeval polymerization thermodynamically more favourable by attracting reactants and arranging them appropriately relative to each other.^[216] The Zn-world theory specifies and differentiates further.^{[215][217]} Hydrothermal fluids rich in H₂S interacting with cold primordial ocean (or Darwin's "warm little pond") water leads to the precipitation of metal sulfide particles. Oceanic vent systems and other hydrothermal systems have a zonal structure reflected in ancient volcanogenic massive sulfide deposits (VMS) of hydrothermal origin. They reach many kilometres in diameter and date back to the Archean Eon. Most abundant are pyrite (FeS₂), chalcopyrite (CuFeS₂), and sphalerite (ZnS), with additions of galena (PbS) and alabandite (MnS). ZnS and MnS have a unique ability to store radiation energy, e.g. provided by UV light. Since during the relevant time window of the origins of replicating molecules the primordial atmospheric pressure was high enough (>100 bar, about 100 atmospheres) to precipitate near the Earth's surface and UV irradiation was 10 to 100 times more intense than now, the unique photosynthetic properties mediated by ZnS provided just the right energy conditions to energize the synthesis of informational and metabolic molecules and the selection of photostable nucleobases.

The Zn-world theory has been further filled out with experimental and theoretical evidence for the ionic constitution of the interior of the first proto-cells before archaea, bacteria and proto-eukaryotes evolved. Archibald Macallum noted the resemblance of organism fluids such as blood, and lymph to seawater;^[218] however, the inorganic composition of all cells differ from that of modern seawater, which led Mulkidjanian and colleagues to reconstruct the "hatcheries" of the first cells combining geochemical analysis with phylogenomic scrutiny of the inorganic ion requirements of universal components of modern cells. The authors conclude that ubiquitous, and by inference primordial, proteins and functional systems show affinity to and functional requirement for K⁺, Zn²⁺, Mn²⁺, and phosphate. Geochemical reconstruction shows that the ionic composition conducive to the origin of cells could not have existed in what we today call marine settings but is compatible with emissions of vapour-dominated zones of what we today call inland geothermal systems. Under the oxygen depleted, CO₂-dominated primordial atmosphere, the chemistry of water condensates and exhalations near geothermal fields would resemble the internal milieu of modern cells. Therefore, the precellular stages of evolution may have taken place in shallow "Darwin ponds" lined with porous silicate minerals mixed with metal sulfides and enriched in K⁺, Zn²⁺, and phosphorus compounds.^{[219][220]}

Deep sea vent hypothesis

The deep sea vent, or alkaline hydrothermal vent, theory for the origin of life on Earth posits that life may have begun at submarine hydrothermal vents,^{[221][222]} William Martin and Michael Russell have suggested "that life evolved in structured iron monosulphide precipitates in a seepage site hydrothermal mound at a redox, pH and temperature gradient between sulphide-rich hydrothermal fluid and iron(II)-containing waters of the Hadean ocean floor. The naturally arising, three-dimensional compartmentation observed within fossilized seepage-site metal sulphide precipitates indicates that these inorganic compartments were the precursors of cell walls and membranes found in free-living prokaryotes. The known capability of FeS and NiS to catalyze the synthesis of the acetyl-methylsulphide from carbon monoxide and methylsulphide, constituents of hydrothermal fluid, indicates that pre-biotic syntheses occurred at the inner surfaces of these metal-sulphide-walled compartments,..."^[223] These form where hydrogen-rich fluids emerge from below the sea floor, as a result of serpentinization of ultra-mafic olivine with seawater and a pH interface with carbon dioxide-rich ocean water. The vents form a sustained chemical energy source derived from redox reactions, in which electron donors, such as molecular hydrogen, react with electron acceptors, such as carbon dioxide (see Iron-sulfur world theory). These are highly exothermic reactions.^{[221][note 2]}



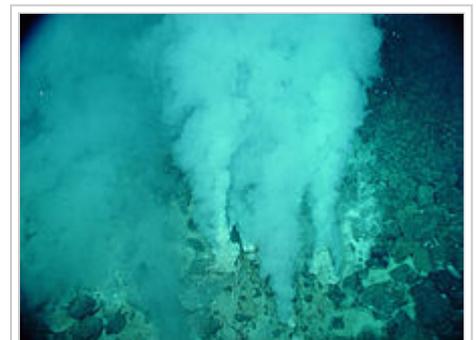
Deep-sea hydrothermal vent or 'black smoker'

Michael Russell demonstrated that alkaline vents created an abiogenic proton motive force (PMF) chemiosmotic gradient,^[223] in which conditions are ideal for an abiogenic hatchery for life. Their microscopic compartments "provide a natural means of concentrating organic molecules," composed of iron-sulfur minerals such as mackinawite, endowed these mineral cells with the catalytic properties envisaged by Wächtershäuser.^[208] This movement of ions across the membrane depends on a combination of two factors:

1. Diffusion force caused by concentration gradient—all particles including ions tend to diffuse from higher concentration to lower.
2. Electrostatic force caused by electrical potential gradient—cations like protons H^+ tend to diffuse down the electrical potential, anions in the opposite direction.

These two gradients taken together can be expressed as an electrochemical gradient, providing energy for abiogenic synthesis. The proton motive force can be described as the measure of the potential energy stored as a combination of proton and voltage gradients across a membrane (differences in proton concentration and electrical potential).

Jack W. Szostak suggested that geothermal activity provides greater opportunities for the origination of life in open lakes where there is a buildup of minerals. In 2010, based on spectral analysis of sea and hot mineral water, Ignat Ignatov and Oleg Mosin demonstrated that life may have predominantly originated in hot mineral water. The hot mineral water that contains bicarbonate and calcium ions has the most optimal range.^[224] This case is similar to the origin of life in hydrothermal vents, but with bicarbonate and calcium ions in hot water. This water has a pH of 9–11 and is possible to have the reactions in seawater. According to Melvin Calvin, certain reactions of condensation-dehydration of amino acids and nucleotides in individual blocks of peptides and nucleic acids can take place in the primary hydrosphere with pH 9-11 at a later evolutionary stage.^[225] Some of these compounds like hydrocyanic acid (HCN) have been proven in the experiments of Miller. This is the environment in which the stromatolites have been created. David Ward of Montana State University described the formation of



White smokers emitting liquid carbon dioxide (CO_2) at the *Champagne* vent, Marianas Trench Marine National Monument

stromatolites in hot mineral water at the Yellowstone National Park. Stromatolites survive in hot mineral water and in proximity to areas with volcanic activity.^[226] Processes have evolved in the sea near geysers of hot mineral water. In 2011, Tadashi Sugawara from the University of Tokyo created a protocell in hot water.^[227]

Experimental research and computer modelling suggest that the surfaces of mineral particles inside hydrothermal vents have catalytic properties similar to those of enzymes and are able to create simple organic molecules, such as methanol (CH₃OH) and formic, acetic and pyruvic acid out of the dissolved CO₂ in the water.^{[228][229]}

The research reported above by William F. Martin in July 2016 supports the thesis that life arose at hydrothermal vents,^{[230][231]} that spontaneous chemistry in the Earth's crust driven by rock–water interactions at disequilibrium thermodynamically underpinned life's origin^{[232][233]} and that the founding lineages of the archaea and bacteria were H₂-dependent autotrophs that used CO₂ as their terminal acceptor in energy metabolism.^[234] Martin suggests, based upon this evidence that LUCA "may have depended heavily on the geothermal energy of the vent to survive".^[235]

Thermosynthesis

Today's bioenergetic process of fermentation is carried out by either the aforementioned citric acid cycle or the Acetyl-CoA pathway, both of which have been connected to the primordial Iron–sulfur world. In a different approach, the thermosynthesis hypothesis considers the bioenergetic process of chemiosmosis, which plays an essential role in cellular respiration and photosynthesis, more basal than fermentation: the ATP synthase enzyme, which sustains chemiosmosis, is proposed as the currently extant enzyme most closely related to the first metabolic process.^{[236][237]}

First, life needed an energy source to bring about the condensation reaction that yielded the peptide bonds of proteins and the phosphodiester bonds of RNA. In a generalization and thermal variation of the binding change mechanism of today's ATP synthase, the "first protein" would have bound substrates (peptides, phosphate, nucleosides, RNA 'monomers') and condensed them to a reaction product that remained bound until after a temperature change it was released by thermal unfolding.

The energy source under the thermosynthesis hypothesis was thermal cycling, the result of suspension of protocells in a convection current, as is plausible in a volcanic hot spring; the convection accounts for the self-organization and dissipative structure required in any origin of life model. The still ubiquitous role of thermal cycling in germination and cell division is considered a relic of primordial thermosynthesis.

By phosphorylating cell membrane lipids, this "first protein" gave a selective advantage to the lipid protocell that contained the protein. This protein also synthesized a library of many proteins, of which only a minute fraction had thermosynthesis capabilities. As proposed by Dyson,^[12] it propagated functionally: it made daughters with similar capabilities, but it did not copy itself. Functioning daughters consisted of different amino acid sequences.

Whereas the Iron–sulfur world identifies a circular pathway as the most simple, the thermosynthesis hypothesis does not even invoke a pathway: ATP synthase's binding change mechanism resembles a physical adsorption process that yields free energy,^[238] rather than a regular enzyme's mechanism, which decreases the free energy. It has been claimed that the emergence of cyclic systems of protein catalysts is implausible.^[239]

Other models

Clay hypothesis

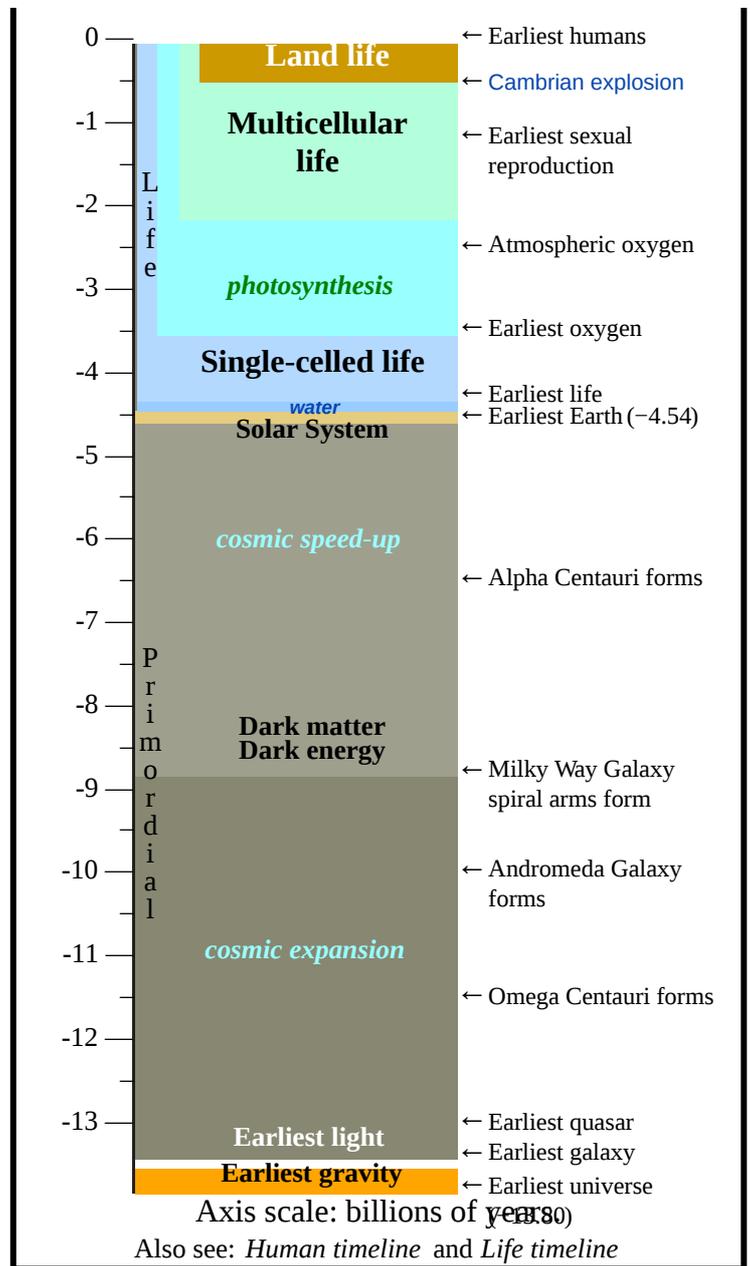
Nature timeline

[view](#) • [discuss](#) •

Montmorillonite, an abundant clay, is a catalyst for the polymerization of RNA and for the formation of membranes from lipids.^[240] A model for the origin of life using clay was forwarded by Alexander Graham Cairns-Smith in 1985 and explored as a plausible mechanism by several scientists.^[241] The clay hypothesis postulates that complex organic molecules arose gradually on pre-existing, non-organic replication surfaces of silicate crystals in solution.

At the Rensselaer Polytechnic Institute, James P. Ferris' studies have also confirmed that clay minerals of montmorillonite catalyze the formation of RNA in aqueous solution, by joining nucleotides to form longer chains.^[242]

In 2007, Bart Kahr from the University of Washington and colleagues reported their experiments that tested the idea that crystals can act as a source of transferable information, using crystals of potassium hydrogen phthalate. "Mother" crystals with imperfections were cleaved and used as seeds to grow "daughter" crystals from solution. They then examined the distribution of imperfections in the new crystals and found that the imperfections in the mother crystals were reproduced in the daughters, but the daughter crystals also had many additional imperfections. For gene-like behaviour to be observed, the quantity of inheritance of these imperfections should have exceeded that of the mutations in the successive generations, but it did not. Thus Kahr concluded that the crystals "were not faithful enough to store and transfer information from one generation to the next."^[243]



Gold's "deep-hot biosphere" model

In the 1970s, Thomas Gold proposed the theory that life first developed not on the surface of the Earth, but several kilometres below the surface. It is claimed that discovery of microbial life below the surface of another body in our Solar System would lend significant credence to this theory. Thomas Gold also asserted that a trickle of food from a deep, unreachable, source is needed for survival because life arising in a puddle of organic material is likely to consume all of its food and become extinct. Gold's theory is that the flow of such food is due to out-gassing of primordial methane from the Earth's mantle; more conventional explanations of the food supply of deep microbes (away from sedimentary carbon compounds) is that the organisms subsist on hydrogen released by an interaction between water and (reduced) iron compounds in rocks.

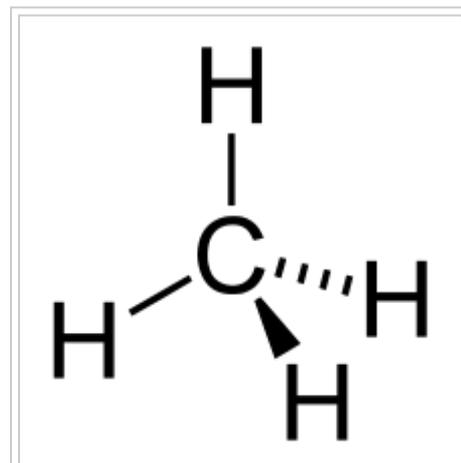
Panspermia

Panspermia is the hypothesis that life exists throughout the Universe, distributed by meteoroids, asteroids, comets,^[244] planetoids,^[245] and, also, by spacecraft in the form of unintended contamination by microorganisms.^{[246][247]}

Panspermia hypothesis does not attempt to explain how life first originated, but merely shifts it to another planet or a comet. The advantage of an extraterrestrial origin of primitive life is that life is not required to have formed on each planet it occurs on, but rather in a single location, and then spread about the galaxy to other star systems via cometary and/or meteorite impact.^[248] Evidence to support the hypothesis is scant, but it finds support in studies of Martian meteorites found in Antarctica and in studies of extremophile microbes' survival in outer space tests.^{[249][250][251][252]} (See also: List of microorganisms tested in outer space.)

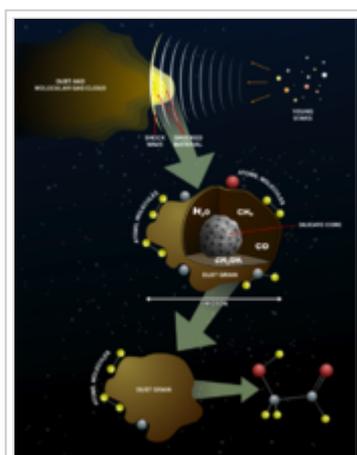
Extraterrestrial organic molecules

An organic compound is any member of a large class of gaseous, liquid, or solid chemicals whose molecules contain carbon. Carbon is the fourth most abundant element in the Universe by mass after hydrogen, helium, and oxygen.^[253] Carbon is abundant in the Sun, stars, comets, and in the atmospheres of most planets.^[254] Organic compounds are relatively common in space, formed by "factories of complex molecular synthesis" which occur in molecular clouds and circumstellar envelopes, and chemically evolve after reactions are initiated mostly by ionizing radiation.^{[19][255][256][257]} Based on computer model studies, the complex organic molecules necessary for life may have formed on dust grains in the protoplanetary disk surrounding the Sun before the formation of the Earth.^[115] According to the computer studies, this same process may also occur around other stars that acquire planets.^[115]



Methane is one of the simplest organic compounds

Observations suggest that the majority of organic compounds introduced on Earth by interstellar dust particles are considered principal agents in the formation of complex molecules, thanks to their peculiar surface-catalytic activities.^{[258][259]} Studies reported in 2008, based on ¹²C/¹³C isotopic ratios of organic compounds found in the Murchison meteorite, suggested that the RNA component uracil and related molecules, including xanthine, were formed extraterrestrially.^{[260][261]} On 8 August 2011, a report based on NASA studies of meteorites found on Earth was published suggesting DNA components (adenine, guanine and related organic molecules) were made in outer space.^{[258][262][263]} Scientists also found that the cosmic dust permeating the Universe contains complex organics ("amorphous organic solids with a mixed aromatic–aliphatic structure") that could be created naturally, and rapidly, by stars.^{[264][265][266]} Sun Kwok of The University of Hong Kong suggested that these compounds may have been related to the development of life on Earth said that "If this is the case, life on Earth may have had an easier time getting started as these organics can serve as basic ingredients for life."^[264]

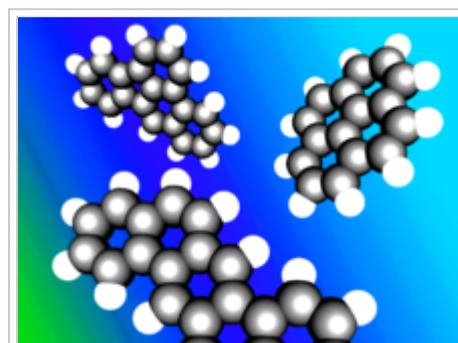


Formation of glycolaldehyde in stardust

Glycolaldehyde, the first example of an interstellar sugar molecule, was detected in the star-forming region near the centre of our galaxy. It was discovered in 2000 by Jes Jørgensen and Jan M. Hollis.^[267] In 2012, Jørgensen's team reported the detection of glycolaldehyde in a distant star system. The molecule was found around the protostellar binary IRAS 16293-2422 400 light years from Earth.^{[268][269][270]} Glycolaldehyde is needed to form RNA, which is similar in function to DNA. These findings suggest that complex organic molecules may form in stellar systems prior to the formation of planets, eventually arriving on young planets early in their formation.^[271] Because sugars are associated with both metabolism and the genetic code, two of the most basic aspects of life, it is thought the discovery of extraterrestrial sugar increases the likelihood that life may exist elsewhere in our galaxy.^[267]

NASA announced in 2009 that scientists had identified another fundamental chemical building block of life in a comet for the first time, glycine, an amino acid, which was detected in material ejected from comet Wild 2 in 2004 and grabbed by NASA's *Stardust* probe. Glycine has been detected in meteorites before. Carl Pilcher, who leads the NASA Astrobiology Institute commented that "The discovery of glycine in a comet supports the idea that the fundamental building blocks of life are prevalent in space, and strengthens the argument that life in the Universe may be common rather than rare."^[272] Comets are encrusted with outer layers of dark material, thought to be a tar-like substance composed of complex organic material formed from simple carbon compounds after reactions initiated mostly by ionizing radiation. It is possible that a rain of material from comets could have brought significant quantities of such complex organic molecules to Earth.^{[273][274][275]} Amino acids which were formed extraterrestrially may also have arrived on Earth via comets.^[47] It is estimated that during the Late Heavy Bombardment, meteorites may have delivered up to five million tons of organic prebiotic elements to Earth per year.^[47]

Polycyclic aromatic hydrocarbons (PAH) are the most common and abundant of the known polyatomic molecules in the observable universe, and are considered a likely constituent of the primordial sea.^{[276][277][278]} In 2010, PAHs, along with fullerenes (or "buckyballs"), have been detected in nebulae.^{[279][280]} In March 2015, NASA scientists reported that, for the first time, complex DNA and RNA organic compounds of life, including uracil, cytosine and thymine, have been formed in the laboratory under outer space conditions, using starting chemicals, such as pyrimidine, found in meteorites. Pyrimidine, like PAHs, the most carbon-rich chemical found in the Universe, may have been formed in red giant stars or in interstellar dust and gas clouds.^[281] A group of Czech scientists reported that all four RNA-bases may be synthesized from formamide in the course of high-energy density events like extraterrestrial impacts.^[282]



An illustration of typical polycyclic aromatic hydrocarbons. Clockwise from top left: benz(e)acephenanthrylene, pyrene and dibenz(ah)anthracene.

Lipid world

The lipid world theory postulates that the first self-replicating object was lipid-like.^{[283][284]} It is known that phospholipids form lipid bilayers in water while under agitation—the same structure as in cell membranes. These molecules were not present on early Earth, but other amphiphilic long-chain molecules also form membranes. Furthermore, these bodies may expand (by insertion of additional lipids), and under excessive expansion may undergo spontaneous splitting which preserves the same size and composition of lipids in the two progenies. The main idea in this theory is that the molecular composition of the lipid bodies is the preliminary way for information storage, and evolution led to the appearance of polymer entities such as RNA or DNA that may store information favourably. Studies on vesicles from potentially prebiotic amphiphiles have so far been limited to systems containing one or two types of amphiphiles. This in contrast to the output of simulated prebiotic chemical reactions, which typically produce very heterogeneous mixtures of compounds.^[163] Within the hypothesis of a lipid bilayer membrane composed of a mixture of various distinct amphiphilic compounds there is the opportunity of a huge number of theoretically possible combinations in the arrangements of these amphiphiles in the membrane. Among all these potential combinations, a specific local arrangement of the membrane would have favoured the constitution of a hypercycle,^{[285][286]} actually a positive feedback composed of two mutual catalysts represented by a membrane site and a specific compound trapped in the vesicle. Such site/compound pairs are transmissible to the daughter vesicles leading to the emergence of distinct lineages of vesicles which would have allowed Darwinian natural selection.^[287]

Polyphosphates

A problem in most scenarios of abiogenesis is that the thermodynamic equilibrium of amino acid versus peptides is in the direction of separate amino acids. What has been missing is some force that drives polymerization. The resolution of this problem may well be in the properties of polyphosphates.^{[288][289]} Polyphosphates are formed by polymerization of ordinary monophosphate ions PO_4^{-3} . Several mechanisms of organic molecule synthesis have been investigated. Polyphosphates cause polymerization of amino acids into peptides. They are also logical precursors in the synthesis of such key biochemical compounds as adenosine triphosphate (ATP). A key issue seems to be that calcium reacts with soluble phosphate to form insoluble calcium phosphate (apatite), so some plausible mechanism must be found to keep calcium ions from causing precipitation of phosphate. There has been much work on this topic over the years, but an interesting new idea is that meteorites may have introduced reactive phosphorus species on the early Earth.^[290]

PAH world hypothesis

Polycyclic aromatic hydrocarbons (PAH) are known to be abundant in the Universe,^{[276][277][278]} including in the interstellar medium, in comets, and in meteorites, and are some of the most complex molecules so far found in space.^[254]

Other sources of complex molecules have been postulated, including extraterrestrial stellar or interstellar origin. For example, from spectral analyses, organic molecules are known to be present in comets and meteorites. In 2004, a team detected traces of PAHs in a nebula.^[291] In 2010, another team also detected PAHs, along with fullerenes, in nebulae.^[279] The use of PAHs has also been proposed as a precursor to the RNA world in the PAH world hypothesis. The Spitzer Space Telescope has detected a star, HH 46-IR, which is forming by a process similar to that by which the Sun formed. In the disk of material surrounding the star, there is a very large range of molecules, including cyanide compounds, hydrocarbons, and carbon monoxide. In September 2012, NASA scientists reported that PAHs, subjected to interstellar medium conditions, are transformed, through hydrogenation, oxygenation and hydroxylation, to more complex organics—"a step along the path toward amino acids and nucleotides, the raw materials of proteins and DNA, respectively."^{[292][293]} Further, as a result of these transformations, the PAHs lose their spectroscopic signature which could be one of the reasons "for the lack of PAH detection in interstellar ice grains, particularly the outer regions of cold, dense clouds or the upper molecular layers of protoplanetary disks."^{[292][293]}

NASA maintains a database for tracking PAHs in the Universe.^{[254][294]} More than 20% of the carbon in the Universe may be associated with PAHs,^{[254][254]} possible starting materials for the formation of life. PAHs seem to have been formed shortly after the Big Bang, are widespread throughout the Universe,^{[276][277][278]} and are associated with new stars and exoplanets.^[254]

Radioactive beach hypothesis

Zachary Adam claims that tidal processes that occurred during a time when the Moon was much closer may have concentrated grains of uranium and other radioactive elements at the high-water mark on primordial beaches, where they may have been responsible for generating life's building blocks.^[295] According to computer models,^[296] a deposit of such radioactive materials could show the same self-sustaining nuclear reaction as that found in the Oklo uranium ore seam in Gabon. Such radioactive beach sand might have provided sufficient energy to generate organic molecules, such as amino acids and sugars from acetonitrile in water. Radioactive monazite material also has released soluble phosphate into the regions between sand-grains, making it biologically "accessible." Thus amino acids, sugars, and soluble phosphates might have been produced simultaneously, according to Adam. Radioactive actinides, left behind in some concentration by the reaction, might have formed part of organometallic complexes. These complexes could have been important early catalysts to living processes.

John Parnell has suggested that such a process could provide part of the "crucible of life" in the early stages of any early wet rocky planet, so long as the planet is large enough to have generated a system of plate tectonics which brings radioactive minerals to the surface. As the early Earth is thought to have had many smaller plates, it might have provided a suitable environment for such processes.^[297]

Thermodynamic dissipation

The 19th-century Austrian physicist Ludwig Boltzmann first recognized that the struggle for existence of living organisms was neither over raw material nor energy, but instead had to do with entropy production derived from the conversion of the solar spectrum into heat by these systems.^[298] Boltzmann thus realized that living systems, like all irreversible processes, were dependent on the dissipation of a generalized chemical potential for their existence. In his book "What is Life", the 20th-century Austrian physicist Erwin Schrödinger^[299] emphasized the importance of Boltzmann's deep insight into the irreversible thermodynamic nature of living systems, suggesting that this was the physics and chemistry behind the origin and evolution of life. However, irreversible processes, and much less living systems, could not be conveniently analyzed under this perspective until Lars Onsager,^[300] and later Ilya Prigogine,^[301] developed an elegant mathematical formalism for treating the "self-organization" of material under a generalized chemical potential. This formalism became known as Classical Irreversible Thermodynamics and Prigogine was awarded the Nobel Prize in Chemistry in 1977 "for his contributions to non-equilibrium thermodynamics, particularly the theory of dissipative structures". The analysis of Prigogine showed that if a system were left to evolve under an imposed external potential, material could spontaneously organize (lower its entropy) forming what he called "dissipative structures" which would increase the dissipation of the externally imposed potential (augment the global entropy production). Non-equilibrium thermodynamics has since been successfully applied to the analysis of living systems, from the biochemical production of ATP^[302] to optimizing bacterial metabolic pathways^[303] to complete ecosystems.^{[304][305][306]}

In his "Thermodynamic Dissipation Theory of the Origin and Evolution of Life",^{[307][308][309]} Karo Michaelian has taken the insight of Boltzmann and the work of Prigogine to its ultimate consequences regarding the origin of life. This theory postulates that the hallmark of the origin and evolution of life is the microscopic dissipative structuring of organic pigments and their proliferation over the entire Earth surface. Present day life augments the entropy production of Earth in its solar environment by dissipating ultraviolet and visible photons into heat through organic pigments in water. This heat then catalyzes a host of secondary dissipative processes such as the water cycle, ocean and wind currents, hurricanes, etc.^{[308][310]} Michaelian argues that if the thermodynamic function of life today is to produce entropy through photon dissipation in organic pigments, then this probably was its function at its very beginnings. It turns out that both RNA and DNA when in water solution are very strong absorbers and extremely rapid dissipaters of ultraviolet light within the 230–290 nm wavelength (UV-C) region, which is a part of the Sun's spectrum that could have penetrated the prebiotic atmosphere.^[311] In fact, not only RNA and DNA, but many fundamental molecules of life (those common to all three domains of life) are also pigments that absorb in the UV-C, and many of these also have a chemical affinity to RNA and DNA.^{[312][313]} Nucleic acids may thus have acted as acceptor molecules to the UV-C photon excited antenna pigment donor molecules by providing an ultrafast channel for dissipation. Michaelian has shown using the formalism of non-linear irreversible thermodynamics that there would have existed during the Archean a thermodynamic imperative to the abiogenic UV-C photochemical synthesis and proliferation of these pigments over the entire Earth surface if they acted as catalysts to augment the dissipation of the solar photons.^[314] By the end of the Archean, with life-induced ozone dissipating UV-C light in the Earth's upper atmosphere, it would have become ever more improbable for a completely new life to emerge that didn't rely on the complex metabolic pathways already existing since now the free energy in the photons arriving at Earth's surface would have been insufficient for direct breaking and remaking of covalent bonds. It has been suggested, however, that such changes in the surface flux of ultraviolet radiation due to geophysical events affecting the atmosphere could have been what promoted the development of complexity in life based on existing metabolic pathways, for example during the Cambrian explosion^[315]

Many salient characteristics of the fundamental molecules of life (those found in all three domains) all point directly to the involvement of UV-C light in the dissipative structuring of incipient life.^[309] Some of the most difficult problems concerning the origin of life, such as enzyme-less replication of RNA and DNA, homochirality of the fundamental molecules, and the origin of information encoding in RNA and DNA, also find an explanation within the same dissipative thermodynamic framework by considering the probable existence of a relation between primordial replication and UV-C photon dissipation. Michaelian suggests that it is erroneous to expect to describe the emergence, proliferation, or even evolution, of life without overwhelming reference to entropy production through the dissipation of a generalized chemical potential, in particular, the prevailing solar photon flux.

Multiple genesis

Different forms of life with variable origin processes may have appeared quasi-simultaneously in the early history of Earth.^[316] The other forms may be extinct (having left distinctive fossils through their different biochemistry—e.g., hypothetical types of biochemistry). It has been proposed that:

The first organisms were self-replicating iron-rich clays which fixed carbon dioxide into oxalic and other dicarboxylic acids. This system of replicating clays and their metabolic phenotype then evolved into the sulfide rich region of the hot spring acquiring the ability to fix nitrogen. Finally phosphate was incorporated into the evolving system which allowed the synthesis of nucleotides and phospholipids. If biosynthesis recapitulates biopoiesis, then the synthesis of amino acids preceded the synthesis of the purine and pyrimidine bases. Furthermore the polymerization of the amino acid thioesters into polypeptides preceded the directed polymerization of amino acid esters by polynucleotides.^[317]

Fluctuating hydrothermal pools on volcanic islands or proto-continents

Armid Mulkidjanian and co-authors think that the marine environments did not provide the ionic balance and composition universally found in cells, as well as of ions required by essential proteins and ribozymes found in virtually all living organisms, especially with respect to K^+/Na^+ ratio, Mn^{2+} , Zn^{2+} and phosphate concentrations. The only known environments that mimic the needed conditions on Earth are found in terrestrial hydrothermal pools fed by steam vents.^[221] Additionally, mineral deposits in these environments under an anoxic atmosphere would have suitable pH (as opposed to current pools in an oxygenated atmosphere), contain precipitates of sulfide minerals that block harmful UV radiation, have wetting/drying cycles that concentrate substrate solutions to concentrations amenable to spontaneous formation of polymers of nucleic acids, and a continual supply of abiotically generated organic molecules, both by chemical reactions in the hydrothermal environment, as well as by exposure to UV light during transport from vents to adjacent pools. Their hypothesized pre-biotic environments are similar to the deep-oceanic vent environments most commonly hypothesized, but add additional components that help explain peculiarities found in reconstructions of the Last Universal Common Ancestor (LUCA) of all living organisms.^[318]

Bruce Damer and David Deamer have come to the conclusion that cell membranes cannot be formed in salty seawater, and must therefore have originated in freshwater. Before the continents formed, the only dry land on Earth would be volcanic islands, where rainwater would form ponds where lipids could form the first stages towards cell membranes. These predecessors of true cells are assumed to have behaved more like a superorganism rather than individual structures, where the porous membranes would house molecules which would leak out and enter other protocells. Only when true cells had evolved would they gradually adapt to saltier environments and enter the ocean.^[319]

Colín-García *et al.* (2016) discuss the advantages and disadvantages of hydrothermal vents as primitive environments.^[221] They mention the exergonic reactions in such systems could have been a source of free energy that promoted chemical reactions, additional to their high mineralogical diversity which implies the induction of important chemical gradients, thus favoring the interaction between electron donors and acceptors. Colín-García *et al.* (2016) also summarize a set of experiments proposed to test the role of hydrothermal vents in prebiotic synthesis.^[221]

Information theory

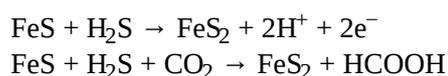
A theory that speaks to the origin of life on Earth and other rocky planets posits life as an information system in which information content grows because of selection. Life must start with minimum possible information, or minimum possible departure from thermodynamic equilibrium, and it requires thermodynamically free energy accessible by means of its information content. The most benign circumstances, minimum entropy variations with abundant free energy, suggest the pore space in the first few kilometres of the surface. Free energy is derived from the condensed products of the chemical reactions taking place in the cooling nebula.^[320]

See also

- Anthropic principle
- Artificial cell
- Artificial life
- Astrochemistry
- Biological immortality
- Common descent
- Emergence
- Entropy and life
- Formamide-based prebiotic chemistry
- GADV protein world
- Mediocrity principle
- Mycoplasma laboratorium
- Nexus for Exoplanet System Science
- Noogenesis
- Planetary habitability
- Rare Earth hypothesis
- Shadow biosphere
- Stromatolite
- Carbon Mineral Challenge

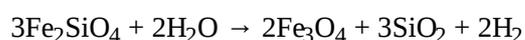
Notes

1. The reactions are:



2. The reactions are:

Reaction 1: Fayalite + water → magnetite + aqueous silica + hydrogen



Reaction 2: Forsterite + aqueous silica → serpentine



Reaction 3: Forsterite + water → serpentine + brucite



Reaction 3 describes the hydration of olivine with water only to yield serpentine and Mg(OH)₂ (brucite). Serpentine is stable at high pH in the presence of brucite like calcium silicate hydrate (C-S-H) phases formed along with portlandite (Ca(OH)₂) in hardened Portland cement paste after the hydration of belite (Ca₂SiO₄), the artificial calcium equivalent of forsterite. Analogy of reaction 3 with belite hydration in ordinary Portland cement



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Absorbance

From Wikipedia, the free encyclopedia

"Optical density" redirects here. "Optical density" can also refer to index of refraction.^[1] For use of the term "optical density" in molecular biology, see Nucleic acid quantitation.

In chemistry, **absorbance** or **decadic absorbance** is the *common logarithm* of the ratio of incident to *transmitted* radiant power through a material, and **spectral absorbance** or **spectral decadic absorbance** is the common logarithm of the ratio of incident to *transmitted* spectral radiant power through a material.^[2] Absorbance is dimensionless, and in particular is not a length, though it is a monotonically increasing function of path length, and approaches zero as the path length approaches zero. The use of the term "optical density" for absorbance is discouraged.^[2] In physics, a closely related quantity called "optical depth" is used instead of absorbance: the *natural logarithm* of the ratio of incident to *transmitted* radiant power through a material. The optical depth equals the absorbance times $\ln(10)$.

The term absorption refers to the physical process of absorbing light, while absorbance does not always measure absorption: it measures attenuation (of transmitted radiant power). Attenuation can be caused by absorption, but also reflection, scattering, and other physical processes.

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Mathematical definitions

Absorbance

Absorbance of a material, denoted A , is given by^[2]

$$A = \log_{10} \left(\frac{\Phi_e^i}{\Phi_e^t} \right) = -\log_{10} T,$$

where

- Φ_e^t is the radiant flux *transmitted* by that material;
- Φ_e^i is the radiant flux received by that material;
- T is the transmittance of that material.

Absorbance is related to optical depth by

$$A = \frac{\tau}{\ln 10},$$

where τ is the optical depth.

Spectral absorbance

Spectral absorbance in frequency and **spectral absorbance in wavelength** of a material, denoted A_ν and A_λ respectively, are given by^[2]

$$A_\nu = \log_{10} \left(\frac{\Phi_{e,\nu}^i}{\Phi_{e,\nu}^t} \right) = -\log_{10} T_\nu,$$
$$A_\lambda = \log_{10} \left(\frac{\Phi_{e,\lambda}^i}{\Phi_{e,\lambda}^t} \right) = -\log_{10} T_\lambda,$$

where

- $\Phi_{e,\nu}^t$ is the spectral radiant flux in frequency *transmitted* by that material;
- $\Phi_{e,\nu}^i$ is the spectral radiant flux in frequency received by that material;
- T_ν is the spectral transmittance in frequency of that material;
- $\Phi_{e,\lambda}^t$ is the spectral radiant flux in wavelength *transmitted* by that material;
- $\Phi_{e,\lambda}^i$ is the spectral radiant flux in wavelength received by that material;
- T_λ is the spectral transmittance in wavelength of that material.

Spectral absorbance is related to spectral optical depth by

$$A_\nu = \frac{\tau_\nu}{\ln 10},$$
$$A_\lambda = \frac{\tau_\lambda}{\ln 10},$$

where

- τ_ν is the spectral optical depth in frequency;
- τ_λ is the spectral optical depth in wavelength.

Although absorbance is properly unitless, it is sometimes reported in "arbitrary units", or AU. Many people, including scientific researchers, wrongly state the results from absorbance measurement experiments in terms of these made-up units.^[3]

Relationship with attenuation

Attenuance

Absorbance is a number that measures the *attenuation* of the transmitted radiant power in a material. Attenuation can be caused by the physical process of "absorption", but also reflection, scattering, and other physical processes. Absorbance of a material is approximately equal to its *attenuance* when both the absorbance is much less than 1 and the emittance of that material (not to be confused with radiant exitance or emissivity) is much less than the absorbance. Indeed,

$$\Phi_e^t + \Phi_e^{\text{att}} = \Phi_e^i + \Phi_e^e,$$

where

- Φ_e^t is the radiant power transmitted by that material;
- Φ_e^{att} is the radiant power attenuated by that material;
- Φ_e^i is the radiant power received by that material;
- Φ_e^e is the radiant power emitted by that material,

that is equivalent to

$$T + ATT = 1 + E,$$

where

- $T = \Phi_e^t / \Phi_e^i$ is the transmittance of that material;
- $ATT = \Phi_e^{\text{att}} / \Phi_e^i$ is the *attenuance* of that material;
- $E = \Phi_e^e / \Phi_e^i$ is the emittance of that material,

and according to Beer–Lambert law, $T = 10^{-A}$, so

$$ATT = 1 - 10^{-A} + E \approx A + E \quad \text{if } A \ll 1,$$

and finally

$$ATT \approx A \quad \text{if } E \ll A.$$

Attenuation coefficient

Absorbance of a material is also related to its *decadic attenuation coefficient* by

$$A = \int_0^l a(z) dz,$$

where

- l is the thickness of that material through which the light travels;
- $a(z)$ is the *decadic attenuation coefficient* of that material at z ,

and if $a(z)$ is uniform along the path, the attenuation is said to be a *linear attenuation* and the relation becomes:

$$A = al.$$

Sometimes the relation is given using the *molar attenuation coefficient* of the material, that is its attenuation coefficient divided by its molar concentration:

$$A = \int_0^l \varepsilon c(z) dz,$$

where

- ε is the *molar attenuation coefficient* of that material;
- $c(z)$ is the molar concentration of that material at z ,

and if $c(z)$ is uniform along the path, the relation becomes:

$$A = \epsilon cl.$$

The use of the term "molar absorptivity" for molar attenuation coefficient is discouraged.^[2]

Measurements

Logarithmic vs. directly proportional measurements

The amount of light transmitted through a material diminishes exponentially as it travels through the material, according to the Beer–Lambert law ($A=(\epsilon)l$). Since the absorbance of a sample is measured as a logarithm, it is directly proportional to the thickness of the sample and to the concentration of the absorbing material in the sample. Some other measures related to absorption, such as transmittance, are measured as a simple ratio so they vary exponentially with the thickness and concentration of the material.

Absorbance: $-\log_{10}(\Phi_e^t/\Phi_e^i)$	Transmittance: Φ_e^t/Φ_e^i
0	1
0.1	0.79
0.25	0.56
0.5	0.32
0.75	0.18
0.9	0.13
1	0.1
2	0.01
3	0.001

Instrument measurement range

Any real measuring instrument has a limited range over which it can accurately measure absorbance. An instrument must be calibrated and checked against known standards if the readings are to be trusted. Many instruments will become non-linear (fail to follow the Beer–Lambert law) starting at approximately 2 AU (~1% transmission). It is also difficult to accurately measure very small absorbance values (below 10^{-4}) with commercially available instruments for chemical analysis. In such cases, laser-based absorption techniques can be used, since they have demonstrated detection limits that supersede those obtained by conventional non-laser-based instruments by many orders of magnitude (detections have been demonstrated all the way down to 5×10^{-13}). The theoretical best accuracy for most commercially available non-laser-based instruments is in the range near 1 AU. The path length or concentration should then, when possible, be adjusted to achieve readings near this range.

Method of measurement

Typically, absorbance of a dissolved substance is measured using absorption spectroscopy. This involves shining a light through a solution and recording how much light and what wavelengths were transmitted onto a detector. Using this information, the wavelengths that were absorbed can be determined.^[4] First, measurements on a "blank" are taken using just the solvent for reference purposes. This is so that the absorbance of the solvent is known, and then any change in absorbance when measuring the whole solution is made by just the solute of interest. Then measurements of the solution are taken. The transmitted spectral radiant flux that makes it through the solution sample is measured and compared to the incident spectral radiant flux. As stated above, the spectral absorbance at a given wavelength is

$$A_{\lambda} = \log_{10} \left(\frac{\Phi_{e,\lambda}^i}{\Phi_{e,\lambda}^t} \right).$$

The absorbance spectrum is plotted on a graph of absorbance vs. wavelength.^[5]

A UV-Vis spectrophotometer will do all this automatically. To use this machine, solutions are placed in a small cuvette and inserted into the holder. The machine is controlled through a computer and, once you "blank" it, will automatically display the absorbance plotted against wavelength. Getting the absorbance spectrum of a solution is useful for determining the concentration of that solution using the Beer–Lambert law and is used in HPLC.

Shade number

Some filters, notably welding glass, are rated by shade number, which is 7/3 times the absorbance plus one:^[6]

$$SN = \frac{7}{3}A + 1,$$

or

$$SN = \frac{7}{3}(-\log_{10} T) + 1,$$

where *SN* is the shade number.

So, if the filter has 0.1% transmittance (0.001 transmittance, which is 3 absorbance units) the shade number would be 8.

See also

- Absorptance
- Tunable Diode Laser Absorption Spectroscopy (TDLAS)
- Densitometry
- Neutral density filter
- Mathematical descriptions of opacity

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Categories: Spectroscopy | Optical filters | Logarithmic scales of measurement

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Albedo

From Wikipedia, the free encyclopedia

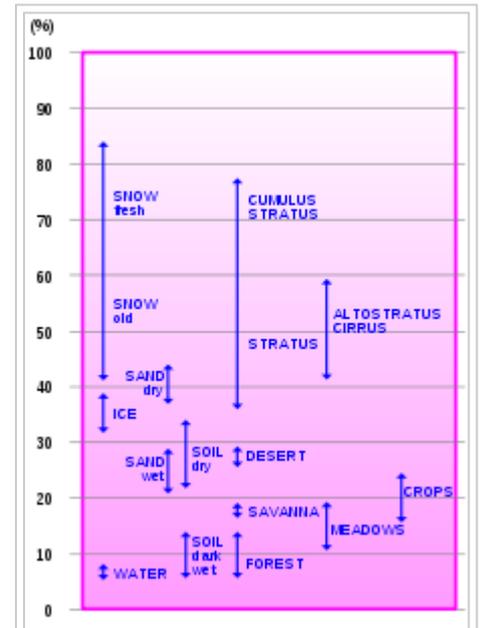
Albedo (/ælˈbiːdoʊ/) is a measure for reflectance or optical brightness (Latin *albedo*, Arabic *albayad*, "whiteness") of a surface. It is dimensionless and measured on a scale from zero (corresponding to a black body that absorbs all incident radiation) to one (corresponding to a white body that reflects all incident radiation).

Surface albedo is defined as the ratio of irradiance reflected to the irradiance received by a surface. The proportion reflected is not only determined by properties of the surface itself, but also by the spectral and angular distribution of solar radiation reaching the Earth's surface.^[1] These factors vary with atmospheric composition, geographic location and time (see Position of the Sun). While bi-hemispherical reflectance is calculated for a single angle of incidence (i.e., for a given position of the sun), albedo is the directional integration of reflectance over all solar angles in a given period. The temporal resolution may range from seconds (as obtained from flux measurements) to daily, monthly or annual averages.

Unless given for a specific wavelength (spectral albedo), albedo refers to the entire spectrum of solar radiation.^[2] Due to measurement constraints, it is often given for the spectrum in which most solar energy reaches the surface (approximately between 0.3 and 3 μm). This spectrum includes visible light (0.39-0.7 μm), which explains why surfaces with a low albedo appear dark (e.g., trees absorb most radiation), whereas surfaces with a high albedo appear bright (e.g., snow reflects most radiation).

Albedo is an important concept in climatology, astronomy, and environmental management (e.g., as part of the Leadership in Energy and Environmental Design (LEED) program for sustainable rating of buildings). The average albedo of the Earth at the top of the atmosphere, its *planetary albedo*, is 30 to 35% because of cloud cover, but widely varies locally across the surface because of different geological and environmental features.^[3]

The term albedo was introduced into optics by Johann Heinrich Lambert in his 1760 work *Photometria*.



Percentage of diffusely reflected sunlight in relation to various surface conditions

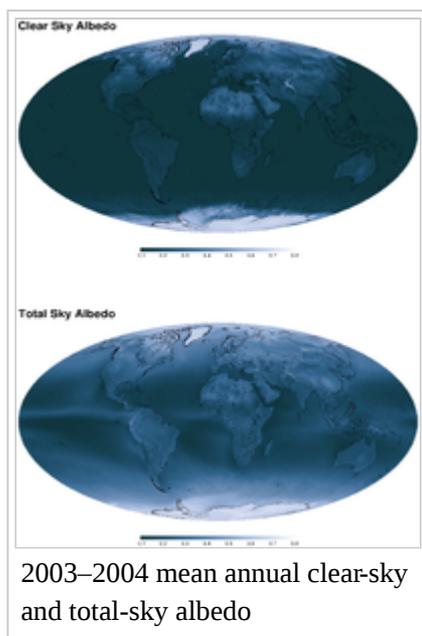
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Terrestrial albedo

Any albedo in visible light falls within a range of about 0.9 for fresh snow to about 0.04 for charcoal, one of the darkest substances. Deeply shadowed cavities can achieve an effective albedo approaching the zero of a black body. When seen from a distance, the ocean surface has a low albedo, as do most forests, whereas desert areas have some of the highest albedos among landforms. Most land areas are in an albedo range of 0.1 to 0.4.^[10] The average albedo of Earth is about 0.3.^[11] This is far higher than for the ocean primarily because of the contribution of clouds.



Earth's surface albedo is regularly estimated via Earth observation satellite sensors such as NASA's MODIS instruments on board the Terra and Aqua satellites, and the CERES instrument on the Suomi NPP and JPSS. As the amount of reflected radiation is only measured for a single direction by satellite, not all directions, a mathematical model is used to translate a sample set of satellite reflectance measurements into estimates of directional-hemispherical reflectance and bi-hemispherical reflectance (e.g.,^[12]). These calculations are based on the bidirectional reflectance distribution function (BRDF), which describes how the reflectance of a given surface depends on the view angle of the observer and the solar angle. Thereby, the BRDF allows to translate

observations of reflectance into albedo.

Earth's average surface temperature due to its albedo and the greenhouse effect is currently about 15 °C. If Earth were frozen entirely (and hence be more reflective), the average temperature of the planet would drop below −40 °C.^[13] If only the continental land masses became covered by glaciers, the mean temperature of the planet would drop to about 0 °C.^[14] In contrast, if the entire Earth was covered by water — a so-called aquaplanet — the average temperature on the planet would rise to almost 27 °C.^[15]

White-sky and black-sky albedo

For land surfaces, it has been shown that the albedo at a particular solar zenith angle θ_i can be approximated by the proportionate sum of two terms: the directional-hemispherical reflectance at that solar zenith angle, $\bar{\alpha}(\theta_i)$, and the bi-hemispherical reflectance, $\bar{\alpha}$, with $D - 1$ being the proportion of direct radiation from a given solar angle, and D being the proportion of diffuse illumination.

Sample albedos

Surface	Typical albedo
Fresh asphalt	0.04 ^[4]
Open ocean	0.06 ^[5]
Worn asphalt	0.12 ^[4]
Conifer forest (Summer)	0.08, ^[6] 0.09 to 0.15 ^[7]
Deciduous trees	0.15 to 0.18 ^[7]
Bare soil	0.17 ^[8]
Green grass	0.25 ^[8]
Desert sand	0.40 ^[9]
New concrete	0.55 ^[8]
Ocean ice	0.5–0.7 ^[8]
Fresh snow	0.80–0.90 ^[8]

Hence, the actual albedo α (also called blue-sky albedo) can then be given as:

$$\alpha = (1 - D)\bar{\alpha}(\theta_i) + D\bar{\alpha}.$$

Directional-hemispherical reflectance is sometimes referred to as black-sky albedo and bi-hemispherical reflectance as white-sky albedo. These terms are important because they allow the albedo to be calculated for any given illumination conditions from a knowledge of the intrinsic properties of the surface.^[16]

Astronomical albedo

The albedos of planets, satellites and asteroids can be used to infer much about their properties. The study of albedos, their dependence on wavelength, lighting angle ("phase angle"), and variation in time comprises a major part of the astronomical field of photometry. For small and far objects that cannot be resolved by telescopes, much of what we know comes from the study of their albedos. For example, the absolute albedo can indicate the surface ice content of outer Solar System objects, the variation of albedo with phase angle gives information about regolith properties, whereas unusually high radar albedo is indicative of high metal content in asteroids.

Enceladus, a moon of Saturn, has one of the highest known albedos of any body in the Solar System, with 99% of EM radiation reflected. Another notable high-albedo body is Eris, with an albedo of 0.96.^[17] Many small objects in the outer Solar System^[18] and asteroid belt have low albedos down to about 0.05.^[19] A typical comet nucleus has an albedo of 0.04.^[20] Such a dark surface is thought to be indicative of a primitive and heavily space weathered surface containing some organic compounds.

The overall albedo of the Moon is measured to be around 0.136,^[21] but it is strongly directional and non-Lambertian, displaying also a strong opposition effect.^[22] Although such reflectance properties are different from those of any terrestrial terrains, they are typical of the regolith surfaces of airless Solar System bodies.

Two common albedos that are used in astronomy are the (V-band) geometric albedo (measuring brightness when illumination comes from directly behind the observer) and the Bond albedo (measuring total proportion of electromagnetic energy reflected). Their values can differ significantly, which is a common source of confusion.

In detailed studies, the directional reflectance properties of astronomical bodies are often expressed in terms of the five Hapke parameters which semi-empirically describe the variation of albedo with phase angle, including a characterization of the opposition effect of regolith surfaces.

The correlation between astronomical (geometric) albedo, absolute magnitude and diameter is:^[23]

$$A = \left(\frac{1329 \times 10^{-H/5}}{D} \right)^2,$$

where A is the astronomical albedo, D is the diameter in kilometers, and H is the absolute magnitude.

Examples of terrestrial albedo effects

Illumination

Although the albedo–temperature effect is best known in colder, whiter regions on Earth, the maximum albedo is actually found in the tropics where year-round illumination is greater. The maximum is additionally in the northern hemisphere, varying between three and twelve degrees north.^[24] The minima are found in the

subtropical regions of the northern and southern hemispheres, beyond which albedo increases without respect to illumination.^[24]

Insolation effects

The intensity of albedo temperature effects depend on the amount of albedo and the level of local insolation (solar irradiance); high albedo areas in the arctic and antarctic regions are cold due to low insolation, where areas such as the Sahara Desert, which also have a relatively high albedo, will be hotter due to high insolation. Tropical and sub-tropical rainforest areas have low albedo, and are much hotter than their temperate forest counterparts, which have lower insolation. Because insolation plays such a big role in the heating and cooling effects of albedo, high insolation areas like the tropics will tend to show a more pronounced fluctuation in local temperature when local albedo changes.

Arctic regions notably release more heat back into space than what they absorb, effectively cooling the Earth. This has been a concern since arctic ice and snow has been melting at higher rates due to higher temperatures, creating regions in the arctic that are notably darker (being water or ground which is darker color) and reflects less heat back into space. This feedback loop results in a reduced albedo effect.^[25]

Climate and weather

Albedo affects climate by determining how much radiation a planet absorbs.^[26] The uneven heating of Earth from albedo variations between land, ice, or ocean surfaces can drive weather.

Albedo–temperature feedback

When an area's albedo changes due to snowfall, a snow–temperature feedback results. A layer of snowfall increases local albedo, reflecting away sunlight, leading to local cooling. In principle, if no outside temperature change affects this area (e.g., a warm air mass), the raised albedo and lower temperature would maintain the current snow and invite further snowfall, deepening the snow–temperature feedback. However, because local weather is dynamic due to the change of seasons, eventually warm air masses and a more direct angle of sunlight (higher insolation) cause melting. When the melted area reveals surfaces with lower albedo, such as grass or soil, the effect is reversed: the darkening surface lowers albedo, increasing local temperatures, which induces more melting and thus reducing the albedo further, resulting in still more heating.

Snow

Snow albedo is highly variable, ranging from as high as 0.9 for freshly fallen snow, to about 0.4 for melting snow, and as low as 0.2 for dirty snow.^[27] Over Antarctica they average a little more than 0.8. If a marginally snow-covered area warms, snow tends to melt, lowering the albedo, and hence leading to more snowmelt because more radiation is being absorbed by the snowpack (the ice–albedo positive feedback). Cryoconite, powdery windblown dust containing soot, sometimes reduces albedo on glaciers and ice sheets.^[28] Hence, small errors in albedo can lead to large errors in energy estimates, which is why it is important to measure the albedo of snow-covered areas through remote sensing techniques rather than applying a single value over broad regions.

Small-scale effects

Albedo works on a smaller scale, too. In sunlight, dark clothes absorb more heat and light-coloured clothes reflect it better, thus allowing some control over body temperature by exploiting the albedo effect of the colour of external clothing.^[29]

Solar photovoltaic effects

Albedo can affect the electrical energy output of solar photovoltaic devices. For example, the effects of a spectrally responsive albedo are illustrated by the differences between the spectrally weighted albedo of solar photovoltaic technology based on hydrogenated amorphous silicon (a-Si:H) and crystalline silicon (c-Si)-based compared to traditional spectral-integrated albedo predictions. Research showed impacts of over 10%.^[30] More recently, the analysis was extended to the effects of spectral bias due to the specular reflectivity of 22 commonly occurring surface materials (both human-made and natural) and analyzes the albedo effects on the performance of seven photovoltaic materials covering three common photovoltaic system topologies: industrial (solar farms), commercial flat rooftops and residential pitched-roof applications.^[31]

Trees

Because forests generally have a low albedo, (the majority of the ultraviolet and visible spectrum is absorbed through photosynthesis), some scientists have suggested that greater heat absorption by trees could offset some of the carbon benefits of afforestation (or offset the negative climate impacts of deforestation). In the case of evergreen forests with seasonal snow cover albedo reduction may be great enough for deforestation to cause a net cooling effect.^[32] Trees also impact climate in extremely complicated ways through evapotranspiration. The water vapor causes cooling on the land surface, causes heating where it condenses, acts a strong greenhouse gas, and can increase albedo when it condenses into clouds^[33] Scientists generally treat evapotranspiration as a net cooling impact, and the net climate impact of albedo and evapotranspiration changes from deforestation depends greatly on local climate^[34]

In seasonally snow-covered zones, winter albedos of treeless areas are 10% to 50% higher than nearby forested areas because snow does not cover the trees as readily. Deciduous trees have an albedo value of about 0.15 to 0.18 whereas coniferous trees have a value of about 0.09 to 0.15.^[7]

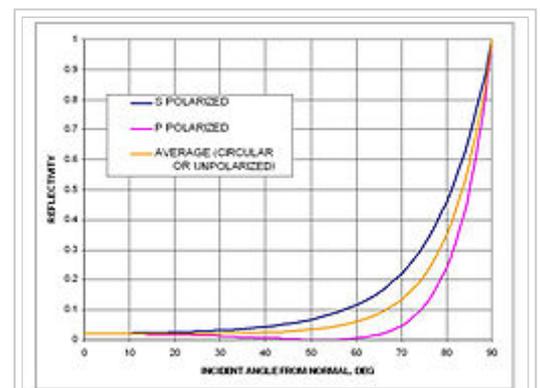
Studies by the Hadley Centre have investigated the relative (generally warming) effect of albedo change and (cooling) effect of carbon sequestration on planting forests. They found that new forests in tropical and midlatitude areas tended to cool; new forests in high latitudes (e.g., Siberia) were neutral or perhaps warming.^[35]

Water

Water reflects light very differently from typical terrestrial materials. The reflectivity of a water surface is calculated using the Fresnel equations (see graph).

At the scale of the wavelength of light even wavy water is always smooth so the light is reflected in a locally specular manner (not diffusely). The glint of light off water is a commonplace effect of this. At small angles of incident light, waviness results in reduced reflectivity because of the steepness of the reflectivity-vs.-incident-angle curve and a locally increased average incident angle.^[36]

Although the reflectivity of water is very low at low and medium angles of incident light, it becomes very high at high angles of incident light such as those that occur on the illuminated side of Earth near the terminator (early morning, late afternoon, and near the poles). However, as mentioned above, waviness causes an appreciable reduction. Because light specularly reflected from water does not usually reach the viewer, water is usually considered to have a very low albedo in spite of its high reflectivity at high angles of incident light.



Reflectivity of smooth water at 20 °C (refractive index=1.333)

Note that white caps on waves look white (and have high albedo) because the water is foamed up, so there are many superimposed bubble surfaces which reflect, adding up their reflectivities. Fresh 'black' ice exhibits Fresnel reflection.

Clouds

Cloud albedo has substantial influence over atmospheric temperatures. Different types of clouds exhibit different reflectivity, theoretically ranging in albedo from a minimum of near 0 to a maximum approaching 0.8. "On any given day, about half of Earth is covered by clouds, which reflect more sunlight than land and water. Clouds keep Earth cool by reflecting sunlight, but they can also serve as blankets to trap warmth."^[37]

Albedo and climate in some areas are affected by artificial clouds, such as those created by the contrails of heavy commercial airliner traffic.^[38] A study following the burning of the Kuwaiti oil fields during Iraqi occupation showed that temperatures under the burning oil fires were as much as 10 °C colder than temperatures several miles away under clear skies.^[39]

Aerosol effects

Aerosols (very fine particles/droplets in the atmosphere) have both direct and indirect effects on Earth's radiative balance. The direct (albedo) effect is generally to cool the planet; the indirect effect (the particles act as cloud condensation nuclei and thereby change cloud properties) is less certain.^[40] As per ^[41] the effects are:

- *Aerosol direct effect.* Aerosols directly scatter and absorb radiation. The scattering of radiation causes atmospheric cooling, whereas absorption can cause atmospheric warming.
- *Aerosol indirect effect.* Aerosols modify the properties of clouds through a subset of the aerosol population called cloud condensation nuclei. Increased nuclei concentrations lead to increased cloud droplet number concentrations, which in turn leads to increased cloud albedo, increased light scattering and radiative cooling (*first indirect effect*), but also leads to reduced precipitation efficiency and increased lifetime of the cloud (*second indirect effect*).

Black carbon

Another albedo-related effect on the climate is from black carbon particles. The size of this effect is difficult to quantify: the Intergovernmental Panel on Climate Change estimates that the global mean radiative forcing for black carbon aerosols from fossil fuels is $+0.2 \text{ W m}^{-2}$, with a range $+0.1$ to $+0.4 \text{ W m}^{-2}$.^[42] Black carbon is a bigger cause of the melting of the polar ice cap in the Arctic than carbon dioxide due to its effect on the albedo.^[43]

Human activities

Human activities (e.g., deforestation, farming, and urbanization) change the albedo of various areas around the globe. However, quantification of this effect on the global scale is difficult.

Other types of albedo

Single-scattering albedo is used to define scattering of electromagnetic waves on small particles. It depends on properties of the material (refractive index); the size of the particle or particles; and the wavelength of the incoming radiation.

See also

- Cool roof
- Daisyworld
- Emissivity
- Global dimming
- Irradiance
- Kirchhoff's law of thermal radiation
- Opposition surge
- Polar see-saw
- Solar radiation management

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External links

- Official Website of Albedo Project
- Global Albedo Project (Center for Clouds, Chemistry, and Climate)
- Albedo – Encyclopedia of Earth
- NASA MODIS BRDF/albedo product site
- Surface albedo derived from Meteosat observations
- A discussion of Lunar albedos
- reflectivity of metals (chart)

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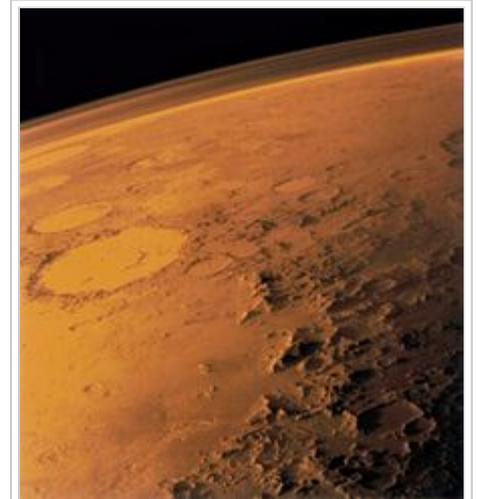
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Atmosphere

From Wikipedia, the free encyclopedia

An **atmosphere** (from Greek *ἀτμός* (*atmos*), meaning 'vapour', and *σφαῖρα* (*sphaira*), meaning 'sphere'^{[1][2]}) is a layer of gases surrounding a planet or other material body, that is held in place by the gravity of that body. An atmosphere is more likely to be retained if the gravity it is subject to is high and the temperature of the atmosphere is low.

The atmosphere of Earth is mostly composed of nitrogen (about 78%), oxygen (about 21%), argon (about 0.9%) with carbon dioxide and other gases in trace amounts. Oxygen is used by most organisms for respiration, nitrogen is fixed by bacteria and lightning to produce ammonia used in the construction of nucleotides and amino acids and carbon dioxide is used by plants, algae and cyanobacteria for photosynthesis. The atmosphere helps protect living organisms from genetic damage by solar ultraviolet radiation, solar wind and cosmic rays. Its current composition is the product of billions of years of biochemical modification of the paleoatmosphere by living organisms.



Mars's thin atmosphere

The term stellar atmosphere describes the outer region of a star, and typically includes the portion starting from the opaque photosphere outwards. Stars with sufficiently low temperatures may form compound molecules in their outer atmosphere.

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Pressure

Atmospheric pressure is the force per unit area that is applied perpendicularly to a surface by the surrounding gas. It is determined by a planet's gravitational force in combination with the total mass of a column of gas above a location. On Earth, units of air pressure are based on the internationally recognized standard atmosphere (atm), which is defined as 101.325 kPa (760 Torr or 14.696 psi). It is measured with a barometer.

The pressure of an atmospheric gas decreases with altitude due to the diminishing mass of gas above. The height at which the pressure from an atmosphere declines by a factor of e (an irrational number with a value of 2.71828..) is called the scale height and is denoted by H . For an atmosphere with a uniform temperature, the scale height is proportional to the temperature and inversely proportional to the product of the mean molecular mass of dry air and the local acceleration of gravity at that location. For such a model atmosphere, the pressure declines exponentially with increasing altitude. However, atmospheres are not uniform in temperature, so the exact determination of the atmospheric pressure at any particular altitude is more complex.

Atmospheric escape

Surface gravity, the force that holds down an atmosphere, differs significantly among the planets. For example, the large gravitational force of the giant planet Jupiter is able to retain light gases such as hydrogen and helium that escape from objects with lower gravity. Secondly, the distance from the Sun determines the energy available to heat atmospheric gas to the point where some fraction of its molecules' thermal motion exceed the planet's escape velocity, allowing those to escape a planet's gravitational grasp. Thus, the distant and cold Titan, Triton, and Pluto are able to retain their atmospheres despite their relatively low gravities. Rogue planets, theoretically, may also retain thick atmospheres.

Since a collection of gas molecules may be moving at a wide range of velocities, there will always be some fast enough to produce a slow leakage of gas into space. Lighter molecules move faster than heavier ones with the same thermal kinetic energy, and so gases of low molecular weight are lost more rapidly than those of high molecular weight. It is thought that Venus and Mars may have lost much of their water when, after being photo dissociated into hydrogen and oxygen by solar ultraviolet, the hydrogen escaped. Earth's magnetic field helps to prevent this, as, normally, the solar wind would greatly enhance the escape of hydrogen. However, over the past 3 billion years Earth may have lost gases through the magnetic polar regions due to auroral activity, including a net 2% of its atmospheric oxygen.^[3]

Other mechanisms that can cause atmosphere depletion are solar wind-induced sputtering, impact erosion, weathering, and sequestration—sometimes referred to as "freezing out"—into the regolith and polar caps.

Terrain

Atmospheres have dramatic effects on the surfaces of rocky bodies. Objects that have no atmosphere, or that have only an exosphere, have terrain that is covered in craters. Without an atmosphere, the planet has no protection from meteors, and all of them collide with the surface and create craters.



A rocky body with a thick atmosphere does not have significant craters on its surface. The friction generated when a meteor enters an atmosphere causes the vast majority of it to burn up before hitting the surface. When meteors do impact, the effects are often erased by the action of wind. As a result, craters are rare on objects with atmospheres.

All objects with atmospheres have wind and weather. Wind erosion is a significant factor in shaping the terrain of rocky planets with atmospheres, and over time can erase the effects of both craters and volcanoes. In addition, since liquids can not exist without pressure, an atmosphere allows liquid to be present at the surface, resulting in lakes, rivers and oceans. Earth and Titan are known to have liquids at their surface and terrain on the planet suggests that Mars had liquid on its surface in the past.

Composition

Initial atmospheric composition is generally related to the chemistry and temperature of the local solar nebula during planetary formation and the subsequent escape of interior gases. The original atmospheres started with the radially local rotating gases that collapsed to the spaced rings that formed the planets. They were then modified over time by various complex factors, resulting in quite different outcomes.

The atmospheres of the planets Venus and Mars are primarily composed of carbon dioxide, with small quantities of nitrogen, argon, oxygen and traces of other gases.

The atmospheric composition on Earth is largely governed by the by-products of the life that it sustains. Dry air from Earth's atmosphere contains 78.08% nitrogen, 20.95% oxygen, 0.93% argon, 0.04% carbon dioxide, and traces of hydrogen, helium, and other "noble" gases (by volume), but generally a variable amount of water vapour is also present, on average about 1% at sea level.

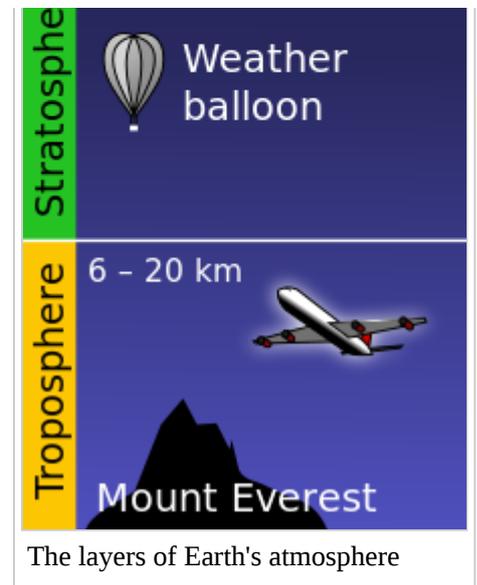
The low temperatures and higher gravity of the Solar System's giant planets—Jupiter, Saturn, Uranus and Neptune—allow them more readily to retain gases with low molecular masses. These planets have hydrogen–helium atmospheres, with trace amounts of more complex compounds.

Two satellites of the outer planets possess significant atmospheres. Titan, a moon of Saturn, and Triton, a moon of Neptune, have atmospheres mainly of nitrogen. When in the part of its orbit closest to the Sun, Pluto has an atmosphere of nitrogen and methane similar to Triton's, but these gases are frozen when it is farther from the Sun.

Other bodies within the Solar System have extremely thin atmospheres not in equilibrium. These include the Moon (sodium gas), Mercury (sodium gas), Europa (oxygen), Io (sulfur), and Enceladus (water vapor).

The first exoplanet whose atmospheric composition was determined is HD 209458b, a gas giant with a close orbit around a star in the constellation Pegasus. Its atmosphere is heated to temperatures over 1,000 K, and is steadily escaping into space. Hydrogen, oxygen, carbon and sulfur have been detected in the planet's inflated atmosphere.^[4]

Structure



Earth's atmospheric gases scatter blue light more than other wavelengths, giving Earth a blue halo when seen from space

Earth

Earth's atmosphere consists of a number of layers, summarised in the diagram above which explains what the layers are, that differ in properties such as composition, temperature and pressure. The lowest layer is the troposphere, which extends from the surface to the bottom of the stratosphere. Three quarters of the atmosphere's mass resides within the troposphere, and is the layer within which the Earth's weather develops. The depth of this layer varies between 17 km at the equator to 7 km at the poles. The stratosphere, extending from the top of the troposphere to the bottom of the mesosphere, contains the ozone layer. The ozone layer ranges in altitude between 15 and 35 km, and is where most of the ultraviolet radiation from the Sun is absorbed. The top of the mesosphere, ranges from 50 to 85 km, and is the layer wherein most meteors burn up. The thermosphere extends from 85 km to the base of the exosphere at 690 km and contains the ionosphere, a region where the atmosphere is ionised by incoming solar radiation. The ionosphere increases in thickness and moves closer to the Earth during daylight and rises at night allowing certain frequencies of radio communication a greater range. The Kármán line, located within the thermosphere at an altitude of 100 km, is commonly used to define the boundary between Earth's atmosphere and outer space. The exosphere begins variously from about 690 to 1,000 km above the surface, where it interacts with the planet's magnetosphere, to space. Each of the layers has a different lapse rate, defining the rate of change in temperature with height.

Others

Other astronomical bodies such as these listed have known atmospheres.

In the Solar System

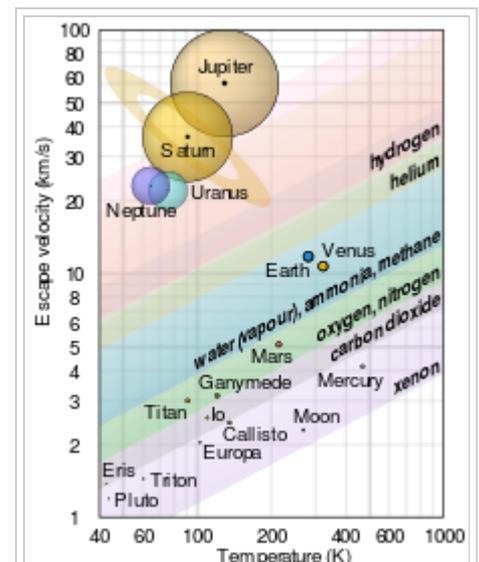
- Atmosphere of the Sun
- Atmosphere of Mercury
- Atmosphere of Venus
- Atmosphere of Earth
 - Atmosphere of the Moon
- Atmosphere of Mars
- Atmosphere of Ceres
- Atmosphere of Jupiter
 - Atmosphere of Io
 - Atmosphere of Callisto
 - Atmosphere of Europa
 - Atmosphere of Ganymede
- Atmosphere of Saturn
 - Atmosphere of Titan
 - Atmosphere of Enceladus
- Atmosphere of Uranus
 - Atmosphere of Titania
- Atmosphere of Neptune
 - Atmosphere of Triton
- Atmosphere of Pluto

Outside the Solar System

- Atmosphere of HD 209458 b

Circulation

The circulation of the atmosphere occurs due to thermal differences when convection becomes a more efficient transporter of heat than thermal radiation. On planets where the primary heat source is solar radiation, excess heat in the tropics is transported to higher latitudes. When a planet generates a significant amount of heat



Graphs of escape velocity against surface temperature of some Solar System objects showing which gases are retained. The objects are drawn to scale, and their data points are at the black dots in the middle.

internally, such as is the case for Jupiter, convection in the atmosphere can transport thermal energy from the higher temperature interior up to the surface.

Importance

From the perspective of a planetary geologist, the atmosphere is an evolutionary agent essential to the shaping of a planet. The wind picks up dust and other particles which when they collide with the terrain erodes the relief and leaves deposits (eolian processes). Frost and precipitations, which depend on the atmospheric composition, also influence the relief. Climate changes can influence a planet's geological history. Conversely, studying surface of Earth leads to an understanding of the atmosphere and climate of a planet—both its present state and its past.

For a meteorologist, the composition of the atmosphere determines the climate and its variations.

For a biologist or paleontologist, the atmospheric composition is closely dependent on the appearance of the life and its evolution.

See also

- Atmometer (evaporimeter)
- Atmospheric pressure
- International Standard Atmosphere
- Kármán
- Sky

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Further reading

- Sanchez-Lavega,, Agustin (2010). *An Introduction to Planetary Atmospheres*. Taylor & Francis. ISBN 978-1-4200-6732-3.

External links

- Properties of atmospheric strata - The flight environment of the atmosphere
- *Atmosphere* - an Open Access journal

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Categories: Atmosphere | Gases | Planetary atmospheres | Planetary science

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Atmosphere of Earth

From Wikipedia, the free encyclopedia

The **atmosphere of Earth** is the layer of gases, commonly known as **air**, that surrounds the planet Earth and is retained by Earth's gravity. The atmosphere of Earth protects life on Earth by absorbing ultraviolet solar radiation, warming the surface through heat retention (greenhouse effect), and reducing temperature extremes between day and night (the diurnal temperature variation).

By volume, dry air contains 78.09% nitrogen, 20.95% oxygen,^[1] 0.93% argon, 0.04% carbon dioxide, and small amounts of other gases. Air also contains a variable amount of water vapor, on average around 1% at sea level, and 0.4% over the entire atmosphere. Air content and atmospheric pressure vary at different layers, and air suitable for use in photosynthesis by terrestrial plants and breathing of terrestrial animals is found only in Earth's troposphere and in artificial atmospheres.

The atmosphere has a mass of about 5.15×10^{18} kg,^[2] three quarters of which is within about 11 km (6.8 mi; 36,000 ft) of the surface. The atmosphere becomes thinner and thinner with increasing altitude, with no definite boundary between the atmosphere and outer space. The Kármán line, at 100 km (62 mi), or 1.57% of Earth's radius, is often used as the border between the atmosphere and outer space. Atmospheric effects become noticeable during atmospheric reentry of spacecraft at an altitude of around 120 km (75 mi). Several layers can be distinguished in the atmosphere, based on characteristics such as temperature and composition.

The study of Earth's atmosphere and its processes is called atmospheric science (aerology). Early pioneers in the field include Léon Teisserenc de Bort and Richard Assmann.^[3]



Blue light is scattered more than other wavelengths by the gases in the atmosphere, giving Earth a blue halo when seen from space onboard *ISS* at a height of 402–424 km (250–263 mi)

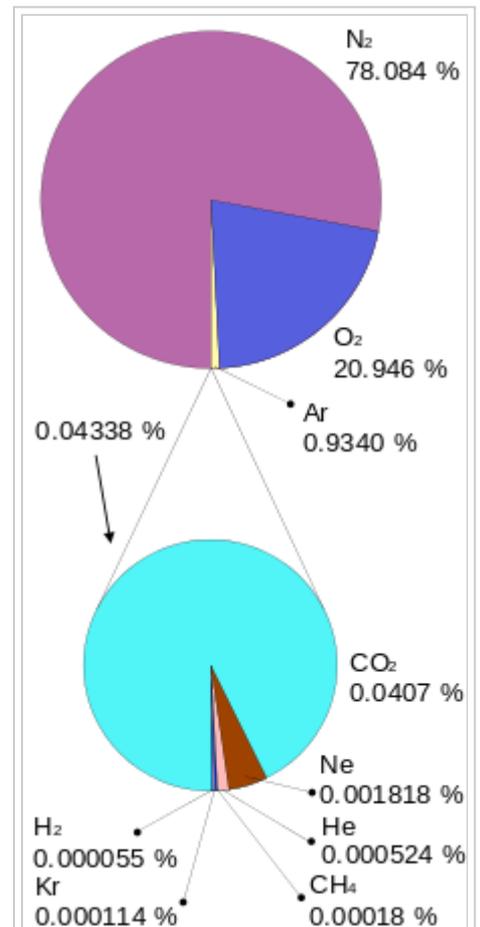
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Composition

The three major constituents of air, and therefore of Earth's atmosphere, are nitrogen, oxygen, and argon. Water vapor accounts for roughly 0.25% of the atmosphere by mass. The concentration of water vapor (a greenhouse gas) varies significantly from around 10 ppm by volume in the coldest portions of the atmosphere to as much as 5% by volume in hot, humid air masses, and concentrations of other atmospheric gases are typically quoted in terms of dry air (without water vapor).^[4] The remaining gases are often referred to as trace gases,^[5] among which are the greenhouse gases, principally carbon dioxide, methane, nitrous oxide, and ozone. Filtered air includes trace amounts of many other chemical compounds. Many substances of natural origin may be present in locally and seasonally variable small amounts as aerosols in an unfiltered air sample, including dust of mineral and organic composition, pollen and spores, sea spray, and volcanic ash. Various industrial pollutants also may be present as gases or aerosols, such as chlorine (elemental or in compounds), fluorine compounds and elemental mercury vapor. Sulfur compounds such as hydrogen sulfide and sulfur dioxide (SO₂) may be derived from natural sources or from industrial air pollution.



Composition of Earth's atmosphere by volume. The lower pie represents the trace gases that together compose about 0.038% of the atmosphere (0.043% with CO₂ at 2014 concentration). The numbers are from a variety of years (mainly 1987, with CO₂ and methane from 2009) and do not represent any single source.

Major constituents of dry air, by volume^[6]

Gas		Volume ^(A)	
Name	Formula	in ppmv ^(B)	in %
Nitrogen	N ₂	780,840	78.084
Oxygen	O ₂	209,460	20.946
Argon	Ar	9,340	0.9340
Carbon dioxide	CO ₂	400	0.04 ^[7]
Neon	Ne	18.18	0.001818
Helium	He	5.24	0.000524
Methane	CH ₄	1.79	0.000179
Not included in above dry atmosphere:			
Water vapor ^(C)	H ₂ O	10–50,000 ^(D)	0.001%–5% ^(D)
notes:			
(A) volume fraction is equal to mole fraction for ideal gas only, also see volume (thermodynamics)			
(B) ppmv: parts per million by volume			
(C) Water vapor is about 0.25% <i>by mass</i> over full atmosphere			
(D) Water vapor strongly varies locally ^[4]			

The relative concentration of gasses remains constant until about 10,000 m (33,000 ft).^[8]

Structure of the atmosphere

Principal layers

In general, air pressure and density decrease with altitude in the atmosphere. However, temperature has a more complicated profile with altitude, and may remain relatively constant or even increase with altitude in some regions (see the temperature section, below). Because the general pattern of the temperature/altitude profile is constant and measurable by means of instrumented balloon soundings, the temperature behavior provides a useful metric to distinguish atmospheric layers. In this way, Earth's atmosphere can be divided (called atmospheric stratification) into five main layers. Excluding the exosphere, the atmosphere has four primary layers, which are the troposphere, stratosphere, mesosphere, and thermosphere.^[9] From highest to lowest, the five main layers are:

- Exosphere: 700 to 10,000 km (440 to 6,200 miles)
- Thermosphere: 80 to 700 km (50 to 440 miles)^[10]
- Mesosphere: 50 to 80 km (31 to 50 miles)
- Stratosphere: 12 to 50 km (7 to 31 miles)
- Troposphere: 0 to 12 km (0 to 7 miles)^[11]

Exosphere

The exosphere is the outermost layer of Earth's atmosphere (i.e. the upper limit of the atmosphere). It extends from the exobase, which is located at the top of the thermosphere at an altitude of about 700 km above sea level, to about 10,000 km (6,200 mi;



33,000,000 ft) where it merges into the solar wind.

This layer is mainly composed of extremely low densities of hydrogen, helium and several heavier molecules including nitrogen, oxygen and carbon dioxide closer to the exobase. The atoms and molecules are so far apart that they can travel hundreds of kilometers without colliding with one another. Thus, the exosphere no longer behaves like a gas, and the particles constantly escape into space. These free-moving particles follow ballistic trajectories and may migrate in and out of the magnetosphere or the solar wind.

The exosphere is located too far above Earth for any meteorological phenomena to be possible. However, the aurora borealis and aurora australis sometimes occur in the lower part of the exosphere, where they overlap into the thermosphere. The exosphere contains most of the satellites orbiting Earth.

Thermosphere

The thermosphere is the second-highest layer of Earth's atmosphere. It extends from the mesopause (which separates it from the mesosphere) at an altitude of about 80 km (50 mi; 260,000 ft) up to the thermopause at an altitude range of 500–1000 km (310–620 mi; 1,600,000–3,300,000 ft). The height of the thermopause varies considerably due to changes in solar activity.^[10] Because the thermopause lies at the lower boundary of the exosphere, it is also referred to as the exobase. The lower part of the thermosphere, from 80 to 550 kilometres (50 to 342 mi) above Earth's surface, contains the ionosphere.

The temperature of the thermosphere gradually increases with height.

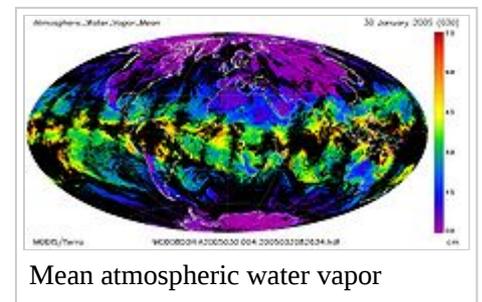
Unlike the stratosphere beneath it, wherein a temperature inversion is due to the absorption of radiation by ozone, the inversion in the thermosphere occurs due to the extremely low density of its molecules. The temperature of this layer can rise as high as 1500 °C (2700 °F), though the gas molecules are so far apart that its temperature in the usual sense is not very meaningful. The air is so rarefied that an individual molecule (of oxygen, for example) travels an average of 1 kilometre (0.62 mi; 3300 ft) between collisions with other molecules.^[12]

Although the thermosphere has a high proportion of molecules with high energy, it would not feel hot to a human in direct contact, because its density is too low to conduct a significant amount of energy to or from the skin.

This layer is completely cloudless and free of water vapor. However, non-hydrometeorological phenomena such as the aurora borealis and aurora australis are occasionally seen in the



Diagram of Earth's atmosphere (layers to scale). Distance from the surface to the top of the stratosphere is just under 1% of Earth's radius.



Mean atmospheric water vapor

- ▼ **Earth's atmosphere** Lower 4 layers of the atmosphere in 3 dimensions as seen diagonally from above the exobase. Layers drawn to scale, objects within the layers are not to scale. Aurorae shown here at the bottom of the thermosphere can actually form at any altitude in this atmospheric layer

thermosphere. The International Space Station orbits in this layer, between 350 and 420 km (220 and 260 mi).

Mesosphere

The mesosphere is the third highest layer of Earth's atmosphere, occupying the region above the stratosphere and below the thermosphere. It extends from the stratopause at an altitude of about 50 km (31 mi; 160,000 ft) to the mesopause at 80–85 km (50–53 mi; 260,000–280,000 ft) above sea level.

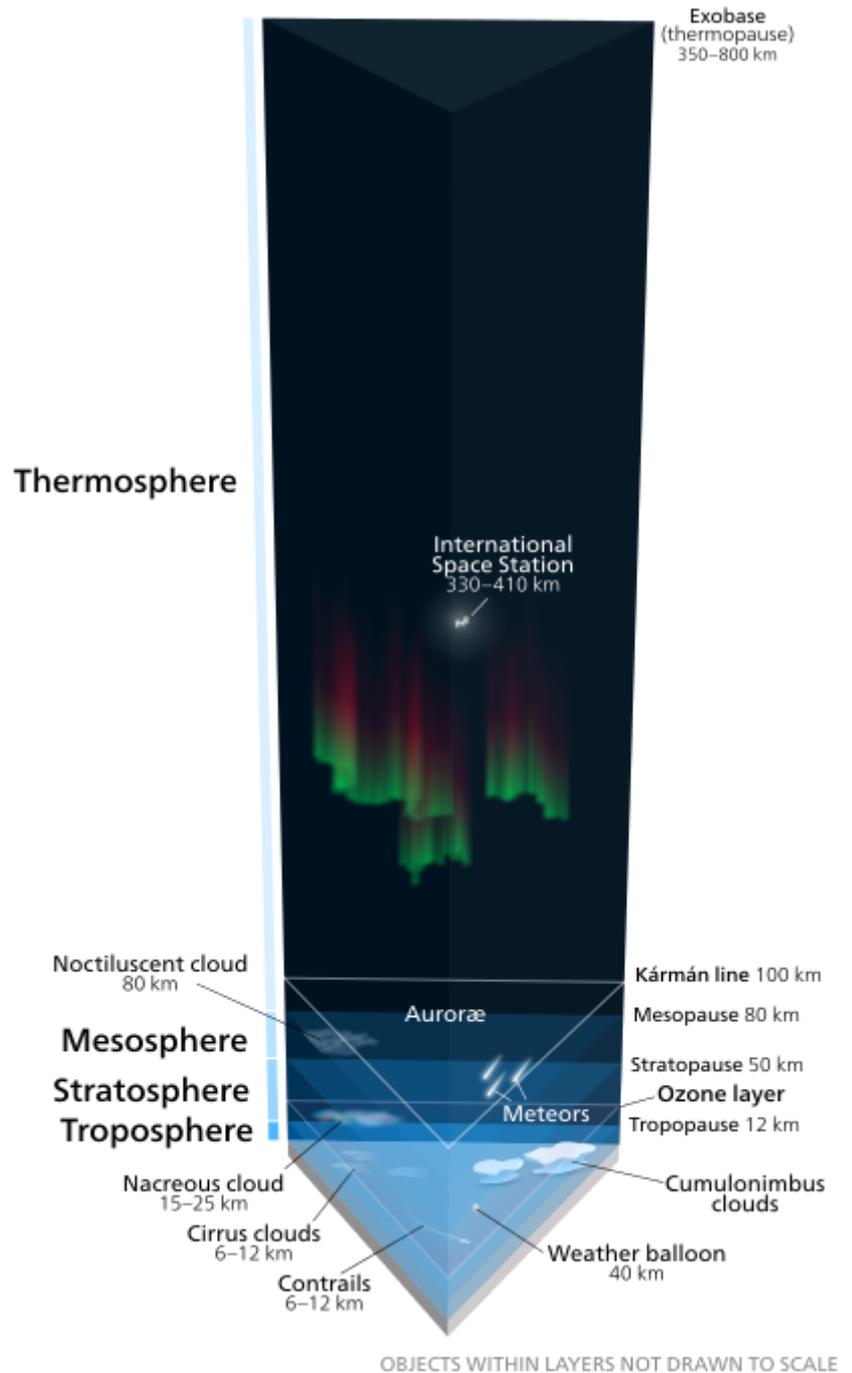
Temperatures drop with increasing altitude to the mesopause that marks the top of this middle layer of the atmosphere. It is the coldest place on Earth and has an average temperature around $-85\text{ }^{\circ}\text{C}$ ($-120\text{ }^{\circ}\text{F}$; 190 K).^{[13][14]}

Just below the mesopause, the air is so cold that even the very scarce water vapor at this altitude can be sublimated into polar-mesospheric noctilucent clouds. These are the highest clouds in the atmosphere and may be visible to the naked eye if sunlight reflects off them about an hour or two after sunset or a similar length of time before sunrise. They are most readily visible when the Sun is around 4 to 16 degrees below the horizon. A type of lightning referred to as either sprites or ELVES occasionally forms far above tropospheric thunderclouds. The mesosphere is also the layer where most meteors burn up upon atmospheric entrance. It is too high above Earth to be accessible to jet-powered aircraft and balloons, and too low to permit orbital spacecraft. The mesosphere is mainly accessed by sounding rockets and rocket-powered aircraft.

Stratosphere

The stratosphere is the second-lowest layer of Earth's atmosphere. It lies above the troposphere and is separated from it by the tropopause. This layer extends from the top of the troposphere at roughly 12 km (7.5 mi; 39,000 ft) above Earth's surface to the stratopause at an altitude of about 50 to 55 km (31 to 34 mi; 164,000 to 180,000 ft).

The atmospheric pressure at the top of the stratosphere is roughly 1/1000 the pressure at sea level. It contains the ozone layer, which is the part of Earth's atmosphere that contains relatively high concentrations of that gas. The stratosphere defines a layer in which temperatures rise with increasing altitude. This rise in temperature is



caused by the absorption of ultraviolet radiation (UV) radiation from the Sun by the ozone layer, which restricts turbulence and mixing. Although the temperature may be $-60\text{ }^{\circ}\text{C}$ ($-76\text{ }^{\circ}\text{F}$; 210 K) at the tropopause, the top of the stratosphere is much warmer, and may be near $0\text{ }^{\circ}\text{C}$.^[15]

The stratospheric temperature profile creates very stable atmospheric conditions, so the stratosphere lacks the weather-producing air turbulence that is so prevalent in the troposphere. Consequently, the stratosphere is almost completely free of clouds and other forms of weather. However, polar stratospheric or nacreous clouds are occasionally seen in the lower part of this layer of the atmosphere where the air is coldest. The stratosphere is the highest layer that can be accessed by jet-powered aircraft.

Troposphere

The troposphere is the lowest layer of Earth's atmosphere. It extends from Earth's surface to an average height of about 12 km, although this altitude actually varies from about 9 km (30,000 ft) at the poles to 17 km (56,000 ft) at the equator,^[11] with some variation due to weather. The troposphere is bounded above by the tropopause, a boundary marked in most places by a temperature inversion (i.e. a layer of relatively warm air above a colder one), and in others by a zone which is isothermal with height.^{[16][17]}

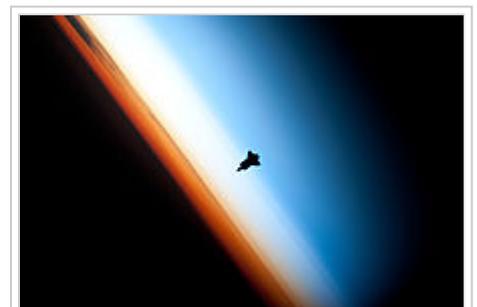
Although variations do occur, the temperature usually declines with increasing altitude in the troposphere because the troposphere is mostly heated through energy transfer from the surface. Thus, the lowest part of the troposphere (i.e. Earth's surface) is typically the warmest section of the troposphere. This promotes vertical mixing (hence the origin of its name in the Greek word *τρόπος*, *tropos*, meaning "turn"). The troposphere contains roughly 80% of the mass of Earth's atmosphere.^[18] The troposphere is denser than all its overlying atmospheric layers because a larger atmospheric weight sits on top of the troposphere and causes it to be most severely compressed. Fifty percent of the total mass of the atmosphere is located in the lower 5.6 km (18,000 ft) of the troposphere.

Nearly all atmospheric water vapor or moisture is found in the troposphere, so it is the layer where most of Earth's weather takes place. It has basically all the weather-associated cloud genus types generated by active wind circulation, although very tall cumulonimbus thunder clouds can penetrate the tropopause from below and rise into the lower part of the stratosphere. Most conventional aviation activity takes place in the troposphere, and it is the only layer that can be accessed by propeller-driven aircraft.

Other layers

Within the five principal layers that are largely determined by temperature, several secondary layers may be distinguished by other properties:

- The ozone layer is contained within the stratosphere. In this layer ozone concentrations are about 2 to 8 parts per million, which is much higher than in the lower atmosphere but still very small compared to the main components of the atmosphere. It is mainly located in the lower portion of the stratosphere from about 15–35 km (9.3–21.7 mi; 49,000–115,000 ft), though the thickness varies seasonally and geographically. About 90% of the ozone in Earth's atmosphere is contained in the stratosphere.
- The ionosphere is a region of the atmosphere that is ionized by solar radiation. It is responsible for auroras. During daytime hours, it stretches from 50 to 1,000 km (31 to 621 mi; 160,000 to 3,280,000 ft) and includes the mesosphere, thermosphere, and parts of the exosphere. However, ionization in the mesosphere largely ceases during the night, so auroras are normally seen only in the thermosphere and lower exosphere. The ionosphere forms the inner edge of the magnetosphere. It has practical importance because it influences, for example, radio propagation on Earth.



Space Shuttle *Endeavour* orbiting in the thermosphere. Because of the angle of the photo, it appears to straddle the stratosphere and mesosphere that actually lie more than 250 km below. The orange layer is the troposphere, which gives way to the whitish stratosphere and then the blue mesosphere.^[19]

- The homosphere and heterosphere are defined by whether the atmospheric gases are well mixed. The surface-based homosphere includes the troposphere, stratosphere, mesosphere, and the lowest part of the thermosphere, where the chemical composition of the atmosphere does not depend on molecular weight because the gases are mixed by turbulence.^[20] This relatively homogeneous layer ends at the *turbopause* found at about 100 km (62 mi; 330,000 ft), the very edge of space itself as accepted by the FAI, which places it about 20 km (12 mi; 66,000 ft) above the mesopause.

Above this altitude lies the heterosphere, which includes the exosphere and most of the thermosphere. Here, the chemical composition varies with altitude. This is because the distance that particles can move without colliding with one another is large compared with the size of motions that cause mixing. This allows the gases to stratify by molecular weight, with the heavier ones, such as oxygen and nitrogen, present only near the bottom of the heterosphere. The upper part of the heterosphere is composed almost completely of hydrogen, the lightest element.

- The planetary boundary layer is the part of the troposphere that is closest to Earth's surface and is directly affected by it, mainly through turbulent diffusion. During the day the planetary boundary layer usually is well-mixed, whereas at night it becomes stably stratified with weak or intermittent mixing. The depth of the planetary boundary layer ranges from as little as about 100 metres (330 ft) on clear, calm nights to 3,000 m (9,800 ft) or more during the afternoon in dry regions.

The average temperature of the atmosphere at Earth's surface is 14 °C (57 °F; 287 K)^[21] or 15 °C (59 °F; 288 K),^[22] depending on the reference.^{[23][24][25]}

Physical properties

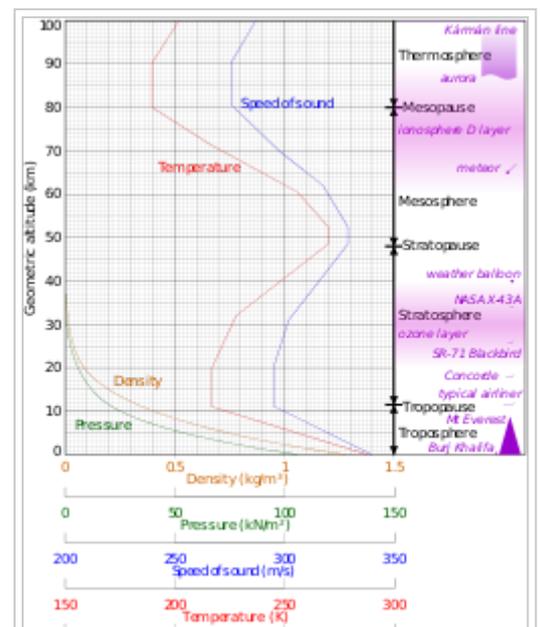
Pressure and thickness

The average atmospheric pressure at sea level is defined by the International Standard Atmosphere as 101325 pascals (760.00 Torr; 14.6959 psi; 760.00 mmHg). This is sometimes referred to as a unit of standard atmospheres (atm). Total atmospheric mass is 5.1480×10^{18} kg (1.135×10^{19} lb),^[27] about 2.5% less than would be inferred from the average sea level pressure and Earth's area of 51007.2 megahectares, this portion being displaced by Earth's mountainous terrain. Atmospheric pressure is the total weight of the air above unit area at the point where the pressure is measured. Thus air pressure varies with location and weather.

If the entire mass of the atmosphere had a uniform density from sea level, it would terminate abruptly at an altitude of 8.50 km (27,900 ft). It actually decreases exponentially with altitude, dropping by half every 5.6 km (18,000 ft) or by a factor of 1/e every 7.64 km (25,100 ft), the average scale height of the atmosphere below 70 km (43 mi; 230,000 ft). However, the atmosphere is more accurately modeled with a customized equation for each layer that takes gradients of temperature, molecular composition, solar radiation and gravity into account.

In summary, the mass of Earth's atmosphere is distributed approximately as follows:^[28]

- 50% is below 5.6 km (18,000 ft).
- 90% is below 16 km (52,000 ft).
- 99.99997% is below 100 km (62 mi; 330,000 ft), the Kármán line. By international convention, this marks the beginning of space where human travelers are considered astronauts.



Comparison of the 1962 US Standard Atmosphere graph of geometric altitude against air density, pressure, the speed of sound and temperature with approximate altitudes of various objects.^[26]

By comparison, the summit of Mt. Everest is at 8,848 m (29,029 ft); commercial airliners typically cruise between 10 km (33,000 ft) and 13 km (43,000 ft) where the thinner air improves fuel economy; weather balloons reach 30.4 km (100,000 ft) and above; and the highest X-15 flight in 1963 reached 108.0 km (354,300 ft).

Even above the Kármán line, significant atmospheric effects such as auroras still occur. Meteors begin to glow in this region, though the larger ones may not burn up until they penetrate more deeply. The various layers of Earth's ionosphere, important to HF radio propagation, begin below 100 km and extend beyond 500 km. By comparison, the International Space Station and Space Shuttle typically orbit at 350–400 km, within the F-layer of the ionosphere where they encounter enough atmospheric drag to require boosts every few months. Depending on solar activity, satellites can experience noticeable atmospheric drag at altitudes as high as 700–800 km.

Temperature and speed of sound

The division of the atmosphere into layers mostly by reference to temperature is discussed above. Temperature decreases with altitude starting at sea level, but variations in this trend begin above 11 km, where the temperature stabilizes through a large vertical distance through the rest of the troposphere. In the stratosphere, starting above about 20 km, the temperature increases with height, due to heating within the ozone layer caused by capture of significant ultraviolet radiation from the Sun by the dioxygen and ozone gas in this region. Still another region of increasing temperature with altitude occurs at very high altitudes, in the aptly-named thermosphere above 90 km.

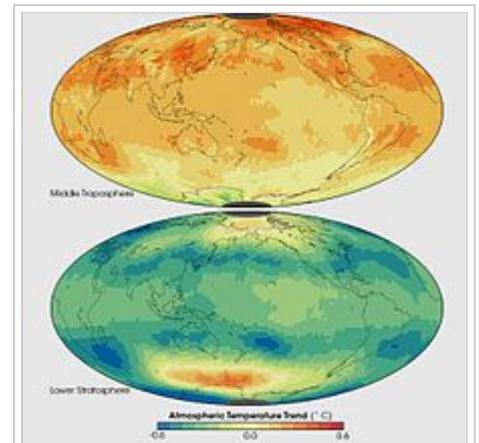
Because in an ideal gas of constant composition the speed of sound depends only on temperature and not on the gas pressure or density, the speed of sound in the atmosphere with altitude takes on the form of the complicated temperature profile (see illustration to the right), and does not mirror altitudinal changes in density or pressure.

Density and mass

The density of air at sea level is about 1.2 kg/m³ (1.2 g/L, 0.0012 g/cm³). Density is not measured directly but is calculated from measurements of temperature, pressure and humidity using the equation of state for air (a form of the ideal gas law). Atmospheric density decreases as the altitude increases. This variation can be approximately modeled using the barometric formula. More sophisticated models are used to predict orbital decay of satellites.

The average mass of the atmosphere is about 5 quadrillion (5×10^{15}) tonnes or 1/1,200,000 the mass of Earth. According to the American National Center for Atmospheric Research, "The total mean mass of the atmosphere is 5.1480×10^{18} kg with an annual range due to water vapor of 1.2 or 1.5×10^{15} kg, depending on whether surface pressure or water vapor data are used; somewhat smaller than the previous estimate. The mean mass of water vapor is estimated as 1.27×10^{16} kg and the dry air mass as $5.1352 \pm 0.0003 \times 10^{18}$ kg."

Optical properties



These images show temperature trends in two thick layers of the atmosphere as measured by a series of satellite-based instruments between January 1979 and December 2005. The measurements were taken by Microwave Sounding Units and Advanced Microwave Sounding Units flying on a series of National Oceanic and Atmospheric Administration (NOAA) weather satellites. The instruments record microwaves emitted from oxygen molecules in the atmosphere. Source:^[29]

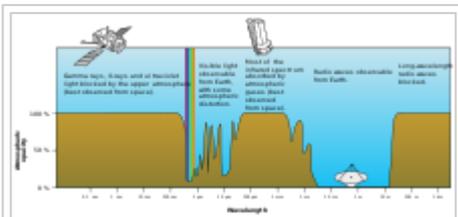
Solar radiation (or sunlight) is the energy Earth receives from the Sun. Earth also emits radiation back into space, but at longer wavelengths that we cannot see. Part of the incoming and emitted radiation is absorbed or reflected by the atmosphere. In May 2017, glints of light, seen as twinkling from an orbiting satellite a million miles away, were found to be reflected light from ice crystals in the atmosphere.^{[30][31]}

Scattering

When light passes through Earth's atmosphere, photons interact with it through *scattering*. If the light does not interact with the atmosphere, it is called *direct radiation* and is what you see if you were to look directly at the Sun. *Indirect radiation* is light that has been scattered in the atmosphere. For example, on an overcast day when you cannot see your shadow there is no direct radiation reaching you, it has all been scattered. As another example, due to a phenomenon called Rayleigh scattering, shorter (blue) wavelengths scatter more easily than longer (red) wavelengths. This is why the sky looks blue; you are seeing scattered blue light. This is also why sunsets are red. Because the Sun is close to the horizon, the Sun's rays pass through more atmosphere than normal to reach your eye. Much of the blue light has been scattered out, leaving the red light in a sunset.

Absorption

Different molecules absorb different wavelengths of radiation. For example, O₂ and O₃ absorb almost all wavelengths shorter than 300 nanometers. Water (H₂O) absorbs many wavelengths above 700 nm. When a molecule absorbs a photon, it increases the energy of the molecule. This heats the atmosphere, but the atmosphere also cools by emitting radiation, as discussed below.



Rough plot of Earth's atmospheric transmittance (or opacity) to various wavelengths of electromagnetic radiation, including visible light.

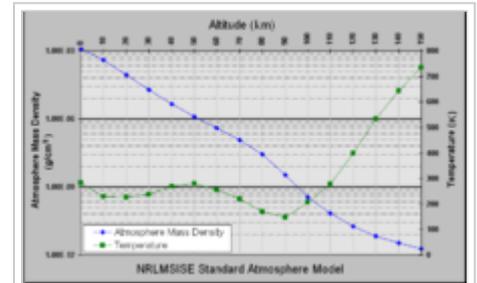
The combined absorption spectra of the gases in the atmosphere leave "windows" of low opacity, allowing the transmission of only certain bands of light. The optical window runs from around 300 nm (ultraviolet-C) up into the range humans can see, the visible spectrum (commonly called light), at roughly 400–700 nm and continues to the infrared to around 1100 nm. There are also infrared and radio windows that transmit some infrared and radio waves at longer wavelengths. For example, the radio window runs from about one centimeter to about eleven-meter waves.

Emission

Emission is the opposite of absorption, it is when an object emits radiation. Objects tend to emit amounts and wavelengths of radiation depending on their "black body" emission curves, therefore hotter objects tend to emit more radiation, with shorter wavelengths. Colder objects emit less radiation, with longer wavelengths. For example, the Sun is approximately 6,000 K (5,730 °C; 10,340 °F), its radiation peaks near 500 nm, and is visible to the human eye. Earth is approximately 290 K (17 °C; 62 °F), so its radiation peaks near 10,000 nm, and is much too long to be visible to humans.

Because of its temperature, the atmosphere emits infrared radiation. For example, on clear nights Earth's surface cools down faster than on cloudy nights. This is because clouds (H₂O) are strong absorbers and emitters of infrared radiation. This is also why it becomes colder at night at higher elevations.

The greenhouse effect is directly related to this absorption and emission effect. Some gases in the atmosphere absorb and emit infrared radiation, but do not interact with sunlight in the visible spectrum. Common examples of these are CO₂ and H₂O.



Temperature and mass density against altitude from the NRLMSISE-00 standard atmosphere model (the eight dotted lines in each "decade" are at the eight cubes 8, 27, 64, ..., 729)

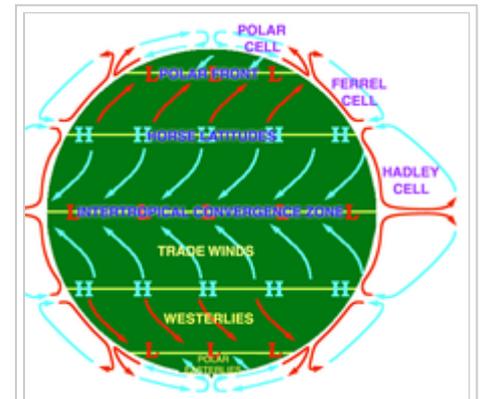
Refractive index

The refractive index of air is close to, but just greater than 1. Systematic variations in refractive index can lead to the bending of light rays over long optical paths. One example is that, under some circumstances, observers onboard ships can see other vessels just over the horizon because light is refracted in the same direction as the curvature of Earth's surface.

The refractive index of air depends on temperature, giving rise to refraction effects when the temperature gradient is large. An example of such effects is the mirage.

Circulation

Atmospheric circulation is the large-scale movement of air through the troposphere, and the means (with ocean circulation) by which heat is distributed around Earth. The large-scale structure of the atmospheric circulation varies from year to year, but the basic structure remains fairly constant because it is determined by Earth's rotation rate and the difference in solar radiation between the equator and poles.



An idealised view of three large circulation cells.

Evolution of Earth's atmosphere

Earliest atmosphere

The first atmosphere consisted of gases in the solar nebula, primarily hydrogen. There were probably simple hydrides such as those now found in the gas giants (Jupiter and Saturn), notably water vapor, methane and ammonia. As the solar nebula dissipated, these gases escaped, partly driven off by the solar wind.^[32]

Second atmosphere

Outgassing from volcanism, supplemented by gases produced during the late heavy bombardment of Earth by huge asteroids, produced the next atmosphere, consisting largely of nitrogen plus carbon dioxide and inert gases.^[32] A major part of carbon-dioxide emissions soon dissolved in water and built up carbonate sediments.

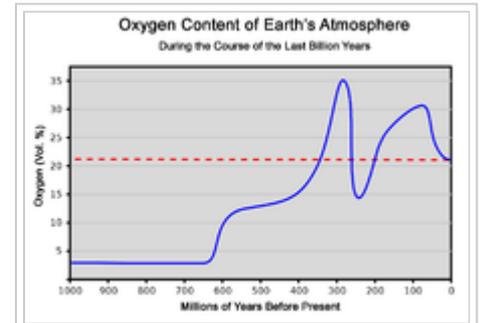
Researchers have found water-related sediments dating from as early as 3.8 billion years ago.^[33] About 3.4 billion years ago, nitrogen formed the major part of the then stable "second atmosphere". An influence of life has to be taken into account rather soon in the history of the atmosphere, because hints of early life-forms appear as early as 3.5 billion years ago.^[34] How Earth at that time maintained a climate warm enough for liquid water and life, if the early Sun put out 30% lower solar radiance than today, is a puzzle known as the "faint young Sun paradox".

The geological record however shows a continually relatively warm surface during the complete early temperature record of Earth - with the exception of one cold glacial phase about 2.4 billion years ago. In the late Archean Eon an oxygen-containing atmosphere began to develop, apparently produced by photosynthesizing cyanobacteria (see Great Oxygenation Event), which have been found as stromatolite fossils from 2.7 billion years ago. The early basic carbon isotopy (isotope ratio proportions) strongly suggests conditions similar to the current, and that the fundamental features of the carbon cycle became established as early as 4 billion years ago.

Ancient sediments in the Gabon dating from between about 2,150 and 2,080 million years ago provide a record of Earth's dynamic oxygenation evolution. These fluctuations in oxygenation were likely driven by the Lomagundi carbon isotope excursion.^[35]

Third atmosphere

The constant re-arrangement of continents by plate tectonics influences the long-term evolution of the atmosphere by transferring carbon dioxide to and from large continental carbonate stores. Free oxygen did not exist in the atmosphere until about 2.4 billion years ago during the Great Oxygenation Event and its appearance is indicated by the end of the banded iron formations. Before this time, any oxygen produced by photosynthesis was consumed by oxidation of reduced materials, notably iron. Molecules of free oxygen did not start to accumulate in the atmosphere until the rate of production of oxygen began to exceed the availability of reducing materials that removed oxygen. This point signifies a shift from a reducing atmosphere to an oxidizing atmosphere. O₂ showed major variations until reaching a steady state of more than 15% by the end of the Precambrian.^[36] The following time span from 541 million years ago to the present day is the Phanerozoic Eon, during the earliest period of which, the Cambrian, oxygen-requiring metazoan life forms began to appear.



Oxygen content of the atmosphere over the last billion years. (2010) This diagram in more detail. See also: Updated figure (2012)

The amount of oxygen in the atmosphere has fluctuated over the last 600 million years, reaching a peak of about 30% around 280 million years ago, significantly higher than today's 21%. Two main processes govern changes in the atmosphere: Plants use carbon dioxide from the atmosphere, releasing oxygen. Breakdown of pyrite and volcanic eruptions release sulfur into the atmosphere, which oxidizes and hence reduces the amount of oxygen in the atmosphere. However, volcanic eruptions also release carbon dioxide, which plants can convert to oxygen. The exact cause of the variation of the amount of oxygen in the atmosphere is not known. Periods with much oxygen in the atmosphere are associated with rapid development of animals. Today's atmosphere contains 21% oxygen, which is high enough for this rapid development of animals.^[37]

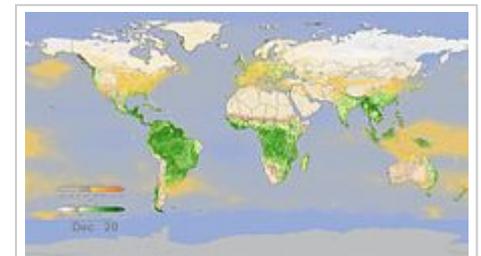
The scientific consensus is that the anthropogenic greenhouse gases currently accumulating in the atmosphere are the main cause of global warming.^[38]

Air pollution

Air pollution is the introduction into the atmosphere of chemicals, particulate matter or biological materials that cause harm or discomfort to organisms.^[39] Stratospheric ozone depletion is caused by air pollution, chiefly from chlorofluorocarbons and other ozone-depleting substances.

Images from space

On October 19, 2015 NASA started a website containing daily images of the full sunlit side of Earth on <http://epic.gsfc.nasa.gov/>. The images are taken from the Deep Space Climate Observatory (DSCOVR) and show Earth as it rotates during a day.^[40]



Play media

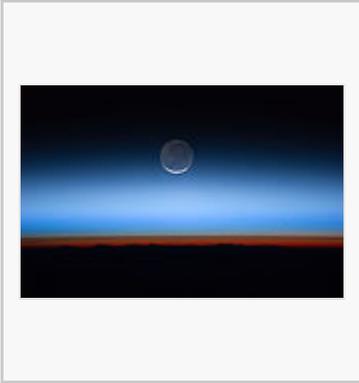
This animation shows the buildup of tropospheric CO₂ in the Northern Hemisphere with a maximum around May. The maximum in the vegetation cycle follows, occurring in the late summer. Following the peak in vegetation, the drawdown of atmospheric CO₂ due to photosynthesis is apparent, particularly over the boreal forests.



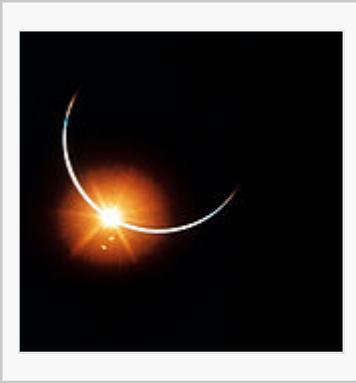
Blue light is scattered more than other wavelengths by the gases in the atmosphere, giving Earth a blue halo when seen from space.

The geomagnetic storms cause beautiful displays of aurora across the atmosphere.

Limb view, of Earth's atmosphere. Colors roughly denote the layers of the atmosphere.



This image shows the Moon at the centre, with the limb of Earth near the bottom transitioning into the orange-colored troposphere. The troposphere ends abruptly at the tropopause, which appears in the image as the sharp boundary between the orange- and blue-colored atmosphere. The silvery-blue noctilucent clouds extend far above Earth's troposphere.



Earth's atmosphere backlit by the Sun in an eclipse observed from deep space onboard Apollo 12 in 1969.

See also

- Aerial perspective
- Air (classical element)
- Air glow
- Airshed
- Atmosphere (for information on atmospheres in general)
- Atmospheric dispersion modeling
- Atmospheric electricity
- Atmospheric models
- Atmospheric Radiation Measurement (ARM) (in the U.S.)
- Atmospheric stratification
- Aviation
- Biosphere

- Carbon dioxide in Earth's atmosphere
- Compressed air
- COSPAR international reference atmosphere (CIRA)
- Environmental impact of aviation
- Global dimming
- Historical temperature record
- Hydrosphere
- Hypermobility (travel)
- Kyoto Protocol
- Leaching (agriculture)
- Lithosphere
- Standard Dry Air
- U.S. Standard Atmosphere
- Warm period
- Water vapor in Earth's atmosphere
- Ozone layer

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External links

- Interactive global map of current atmospheric and ocean surface conditions.

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Atmosphere of Jupiter

From Wikipedia, the free encyclopedia

The **atmosphere of Jupiter** is the largest planetary atmosphere in the Solar System. It is mostly made of molecular hydrogen and helium in roughly solar proportions; other chemical compounds are present only in small amounts and include methane, ammonia, hydrogen sulfide and water. Although water is thought to reside deep in the atmosphere, its directly measured concentration is very low. The nitrogen, sulfur, and noble gas abundances in Jupiter's atmosphere exceed solar values by a factor of about three.^[2]

The atmosphere of Jupiter lacks a clear lower boundary and gradually transitions into the liquid interior of the planet.^[3] From lowest to highest, the atmospheric layers are the troposphere, stratosphere, thermosphere and exosphere. Each layer has characteristic temperature gradients.^[4] The lowest layer, the troposphere, has a complicated system of clouds and hazes, comprising layers of ammonia,

ammonium hydrosulfide and water.^[5] The upper ammonia clouds visible at Jupiter's surface are organized in a dozen zonal bands parallel to the equator and are bounded by powerful zonal atmospheric flows (winds) known as *jets*. The bands alternate in color: the dark bands are called *belts*, while light ones are called *zones*. Zones, which are colder than belts, correspond to upwellings, while belts mark descending air.^[6] The zones' lighter color is believed to result from ammonia ice; what gives the belts their darker colors is uncertain.^[6] The origins of the banded structure and jets are not well understood, though a "shallow model" and a "deep model" exist.^[7]

The Jovian atmosphere shows a wide range of active phenomena, including band instabilities, vortices (cyclones and anticyclones), storms and lightning.^[8] The vortices reveal themselves as large red, white or brown spots (ovals). The largest two spots are the Great Red Spot (GRS)^[9] and Oval BA,^[10] which is also red. These two and most of the other large spots are anticyclonic. Smaller anticyclones tend to be white. Vortices are thought to be relatively shallow structures with depths not exceeding several hundred kilometers. Located in the southern hemisphere, the GRS is the largest known vortex in the Solar System. It could engulf two or three Earths and has existed for at least three hundred years. Oval BA, south of GRS, is a red spot a third the size of GRS that formed in 2000 from the merging of three white ovals.^[11]

Jupiter has powerful storms, often accompanied by lightning strikes. The storms are a result of moist convection in the atmosphere connected to the evaporation and condensation of water. They are sites of strong upward motion of the air, which leads to the formation of bright and dense clouds. The storms form mainly in belt regions. The lightning strikes on Jupiter are hundreds of times more powerful than those seen on Earth, and are assumed to be associated with the water clouds.^[12]



Jupiter's swirling colourful clouds.^[1]

Contents

- 1 Vertical structure
- 2 Chemical composition
- 3 Zones, belts and jets

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- 4 Dynamics
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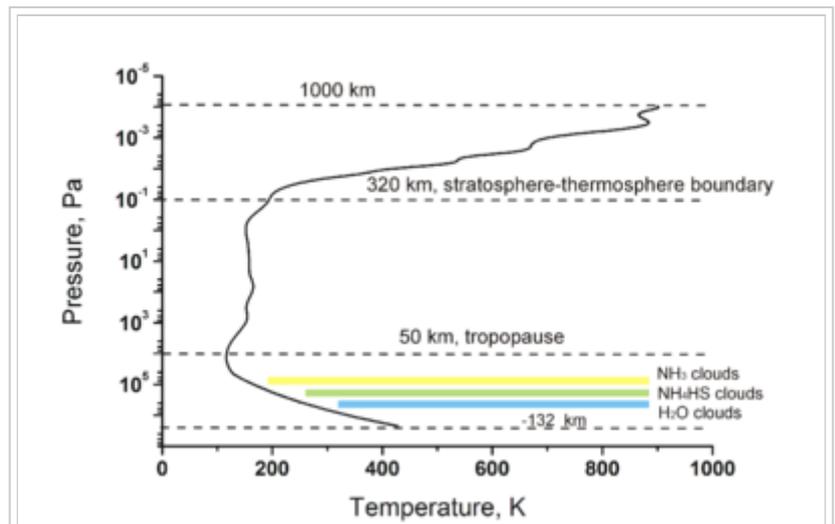
Vertical structure

The atmosphere of Jupiter is classified into four layers, by increasing altitude: the troposphere, stratosphere, thermosphere and exosphere. Unlike the Earth's atmosphere, Jupiter's lacks a mesosphere.^[13] Jupiter does not have a solid surface, and the lowest atmospheric layer, the troposphere, smoothly transitions into the planet's fluid interior.^[3] This is a result of having temperatures and the pressures well above those of the critical points for hydrogen and helium, meaning that there is no sharp boundary between gas and liquid phases. Hydrogen becomes a supercritical fluid at a pressure of around 12 bar.^[3]

Since the lower boundary of the atmosphere is ill-defined, the pressure level of 10 bars, at an altitude of about 90 km below 1 bar with a temperature of around 340 K, is commonly treated as the base of the troposphere.^[4]

In scientific literature, the 1 bar pressure level is usually chosen as a zero point for altitudes—a "surface" of Jupiter.^[3] As with Earth, the top atmospheric layer, the exosphere, does not have a well defined upper boundary.^[14] The density gradually decreases until it smoothly transitions into the interplanetary medium approximately 5,000 km above the "surface".^[15]

The vertical temperature variations in the Jovian atmosphere are similar to those of the atmosphere of Earth. The temperature of the troposphere decreases with height until it reaches a minimum at the tropopause,^[16] which is the boundary between the troposphere and stratosphere. On Jupiter, the tropopause is approximately 50 km above the visible clouds (or 1 bar level), where the pressure and temperature are about 0.1 bar and



Vertical structure of the atmosphere of Jupiter: Note that the temperature drops together with altitude above the tropopause. The *Galileo* atmospheric probe stopped transmitting at a depth of 132 km below the 1 bar "surface" of Jupiter.^[4]

110 K.^{[4][17]} In the stratosphere, the temperatures rise to about 200 K at the transition into the thermosphere, at an altitude and pressure of around 320 km and 1 μbar .^[4] In the thermosphere, temperatures continue to rise, eventually reaching 1000 K at about 1000 km, where pressure is about 1 nbar.^[18]

Jupiter's troposphere contains a complicated cloud structure.^[19] The upper clouds, located in the pressure range 0.6–0.9 bar, are made of ammonia ice.^[20] Below these ammonia ice clouds, denser clouds made of ammonium hydrosulfide or ammonium sulfide (between 1–2 bar) and water (3–7 bar) are thought to exist.^{[21][22]} There are no methane clouds as the temperatures are too high for it to condense.^[19] The water clouds form the densest layer of clouds and have the strongest influence on the dynamics of the atmosphere. This is a result of the higher condensation heat of water and higher water abundance as compared to the ammonia and hydrogen sulfide (oxygen is a more abundant chemical element than either nitrogen or sulfur).^[13] Various tropospheric (at 200–500 mbar) and stratospheric (at 10–100 mbar) haze layers reside above the main cloud layers.^{[21][23]} The latter are made from condensed heavy polycyclic aromatic hydrocarbons or hydrazine, which are generated in the upper stratosphere (1–100 μbar) from methane under the influence of the solar ultraviolet radiation (UV).^[19] The methane abundance relative to molecular hydrogen in the stratosphere is about 10^{-4} ,^[15] while the abundance ratio of other light hydrocarbons, like ethane and acetylene, to molecular hydrogen is about 10^{-6} .^[15]

Jupiter's thermosphere is located at pressures lower than 1 μbar and demonstrates such phenomena as airglow, polar aurorae and X-ray emissions.^[24] Within it lie layers of increased electron and ion density that form the ionosphere.^[15] The high temperatures prevalent in the thermosphere (800–1000 K) have not been fully explained yet;^[18] existing models predict a temperature no higher than about 400 K.^[15] They may be caused by absorption of high-energy solar radiation (UV or X-ray), by heating from the charged particles precipitating from the Jovian magnetosphere, or by dissipation of upward-propagating gravity waves.^[25] The thermosphere and exosphere at the poles and at low latitudes emit X-rays, which were first observed by the Einstein Observatory in 1983.^[26] The energetic particles coming from Jupiter's magnetosphere create bright auroral ovals, which encircle the poles. Unlike their terrestrial analogs, which appear only during magnetic storms, aurorae are permanent features of Jupiter's atmosphere.^[26] The thermosphere was the first place outside the Earth where the trihydrogen cation (H_3^+) was discovered.^[15] This ion emits strongly in the mid-infrared part of the spectrum, at wavelengths between 3 and 5 μm ; this is the main cooling mechanism of the thermosphere.^[24]

Chemical composition

The composition of Jupiter's atmosphere is similar to that of the planet as a whole.^[2] Jupiter's atmosphere is the most comprehensively understood of those of all the gas giants because it was observed directly by the *Galileo* atmospheric probe when it entered the Jovian atmosphere on December 7, 1995.^[27] Other sources of information about Jupiter's atmospheric composition include the *Infrared Space Observatory* (ISO),^[28] the *Galileo* and *Cassini* orbiters,^[29] and Earth-based observations.^[2]

The two main constituents of the Jovian atmosphere are molecular hydrogen (H_2) and helium.^[2] The helium abundance is 0.157 ± 0.0036 relative to molecular hydrogen by number of molecules, and its mass fraction is 0.234 ± 0.005 , which is slightly lower than the Solar System's primordial value.^[2] The reason for this low abundance is not entirely understood, but some of the helium may have condensed into the core of Jupiter.^[20] This condensation is likely to be in the form of helium rain: as hydrogen turns into the metallic state at depths of more than 10,000 km, helium separates from it forming droplets which, being denser than the metallic hydrogen, descend towards the core. This can also explain the severe depletion of neon (see Table), an element that easily dissolves in helium droplets and would be transported in them towards the core as well.^[30]

The atmosphere contains various simple compounds such as water, methane (CH₄), hydrogen sulfide (H₂S), ammonia (NH₃) and phosphine (PH₃).^[2] Their abundances in the deep (below 10 bar) troposphere imply that the atmosphere of Jupiter is enriched in the elements carbon, nitrogen, sulfur and possibly oxygen^[b] by factor of 2–4 relative to the Sun.^{[c][2]} The noble gases argon, krypton and xenon also appear in abundance relative to solar levels (see table), while neon is scarcer.^[2] Other chemical compounds such as arsine (AsH₃) and germane (GeH₄) are present only in trace amounts.^[2] The upper atmosphere of Jupiter contains small amounts of simple hydrocarbons such as ethane, acetylene, and diacetylene, which form from methane under the influence of the solar ultraviolet radiation and charged particles coming from Jupiter's magnetosphere.^[2] The carbon dioxide, carbon monoxide and water present in the upper atmosphere are thought to originate from impacting comets, such as Shoemaker-Levy 9. The water cannot come from the troposphere because the cold tropopause acts like a cold trap, effectively preventing water from rising to the stratosphere (see Vertical structure above).^[2]

Earth- and spacecraft-based measurements have led to improved knowledge of the isotopic ratios in Jupiter's atmosphere. As of July 2003, the accepted value for the deuterium abundance is $2.25 \pm 0.35 \times 10^{-5}$,^[2] which probably represents the primordial value in the protosolar nebula that gave birth to the Solar System.^[28] The ratio of nitrogen isotopes in the Jovian atmosphere, ¹⁵N to ¹⁴N, is 2.3×10^{-3} , a third lower than that in the Earth's atmosphere (3.5×10^{-3}).^[2] The latter discovery is especially significant since the previous theories of Solar System formation considered the terrestrial value for the ratio of nitrogen isotopes to be primordial.^[28]

Zones, belts and jets

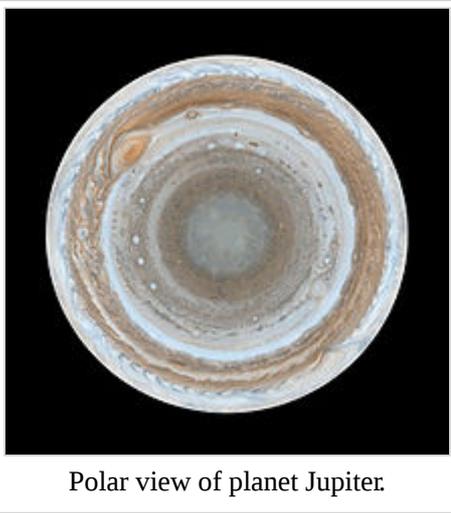
The visible surface of Jupiter is divided into several bands parallel to the equator. There are two types of bands: lightly colored *zones* and relatively dark *belts*.^[6] The wider Equatorial Zone (EZ) extends between latitudes of approximately 7°S to 7°N. Above and below the EZ, the North and South Equatorial belts (NEB and SEB) extend to 18°N and 18°S, respectively. Farther from the equator lie the North and South Tropical zones (NtrZ and STrZ).^[6] The alternating pattern of belts and zones continues until the polar regions at approximately 50 degrees latitude, where their visible appearance becomes somewhat muted.^[31] The basic belt-zone structure probably extends well towards the poles, reaching at least to 80° North or South.^[6]

Elemental abundances relative to hydrogen in Jupiter and Sun^[2]

Element	Sun	Jupiter/Sun
He/H	0.0975	0.807 ± 0.02
Ne/H	1.23×10^{-4}	0.10 ± 0.01
Ar/H	3.62×10^{-6}	2.5 ± 0.5
Kr/H	1.61×10^{-9}	2.7 ± 0.5
Xe/H	1.68×10^{-10}	2.6 ± 0.5
C/H	3.62×10^{-4}	2.9 ± 0.5
N/H	1.12×10^{-4}	3.6 ± 0.5 (8 bar)
		3.2 ± 1.4 (9–12 bar)
O/H	8.51×10^{-4}	0.033 ± 0.015 (12 bar)
		0.19–0.58 (19 bar)
P/H	3.73×10^{-7}	0.82
S/H	1.62×10^{-5}	2.5 ± 0.15

Isotopic ratios in Jupiter and Sun^[2]

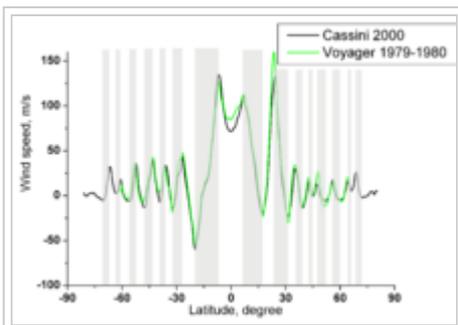
Ratio	Sun	Jupiter
¹³ C/ ¹² C	0.011	0.0108 ± 0.0005
¹⁵ N/ ¹⁴ N	$<2.8 \times 10^{-3}$	$2.3 \pm 0.3 \times 10^{-3}$
		(0.08–2.8 bar)
³⁶ Ar/ ³⁸ Ar	5.77 ± 0.08	5.6 ± 0.25
²⁰ Ne/ ²² Ne	13.81 ± 0.08	13 ± 2
³ He/ ⁴ He	$1.5 \pm 0.3 \times 10^{-4}$	$1.66 \pm 0.05 \times 10^{-4}$
D/H	$3.0 \pm 0.17 \times 10^{-5}$	$2.25 \pm 0.35 \times 10^{-5}$



Polar view of planet Jupiter.

The difference in the appearance between zones and belts is caused by differences in the opacity of the clouds. Ammonia concentration is higher in zones, which leads to the appearance of denser clouds of ammonia ice at higher altitudes, which in turn leads to their lighter color.^[16] On the other hand, in belts clouds are thinner and are located at lower altitudes.^[16] The upper troposphere is colder in zones and warmer in belts.^[6] The exact nature of chemicals that make Jovian zones and bands so colorful is not known, but they may include complicated compounds of sulfur, phosphorus and carbon.^[6]

The Jovian bands are bounded by zonal atmospheric flows (winds), called *jets*. The eastward (prograde) jets are found at the transition from zones to belts (going away from the equator), whereas westward (retrograde) jets mark the transition from belts to zones.^[6] Such flow velocity patterns mean that the zonal winds decrease in belts and increase in zones from the equator to the pole. Therefore, wind shear in belts is cyclonic, while in zones it is anticyclonic.^[22] The EZ is an exception to this rule, showing a strong eastward (prograde) jet and has a local minimum of the wind speed exactly at the equator. The jet speeds are high on Jupiter, reaching more than 100 m/s.^[6] These speeds correspond to ammonia clouds located in the pressure range 0.7–1 bar. The prograde jets are generally more powerful than the retrograde jets.^[6] The vertical extent of jets is not known. They decay over two to three scale heights^[a] above the clouds, while below the cloud level, winds increase slightly and then remain constant down to at least 22 bar—the maximum operational depth reached by the Galileo Probe.^[17]



Zonal wind speeds in the atmosphere of Jupiter

The origin of Jupiter's banded structure is not completely clear, though it may be similar to that driving the Earth's Hadley cells. The simplest interpretation is that zones are sites of atmospheric upwelling, whereas belts are manifestations of downwelling.^[32] When air enriched in ammonia rises in zones, it expands and cools, forming high and dense clouds. In belts, however, the air descends, warming adiabatically as in a convergence zone on Earth, and white ammonia clouds evaporate, revealing lower, darker clouds. The location and width of bands, speed and location of jets on Jupiter are remarkably stable, having changed only slightly between 1980 and 2000. One example of change is a decrease of the speed of the strongest eastward jet located at the boundary between the North Tropical zone and North Temperate belts at

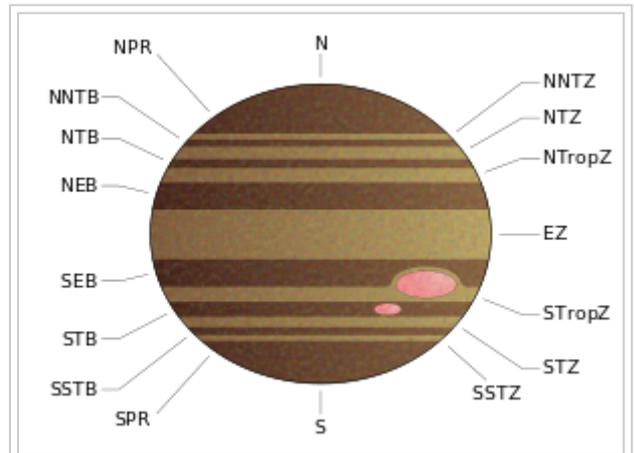
23°N.^{[7][32]} However bands vary in coloration and intensity over time (see below). These variations were first observed in the early seventeenth century.^[33]

Specific bands

The belts and zones that divide Jupiter's atmosphere each have their own names and unique characteristics. They begin below the North and South Polar Regions, which extend from the poles to roughly 40–48° N/S. These bluish-gray regions are usually featureless.^[31]

The North North Temperate Region rarely shows more detail than the polar regions, due to limb darkening, foreshortening, and the general diffuseness of features. However, the North-North Temperate Belt (NNTB) is the northernmost distinct belt, though it occasionally disappears. Disturbances tend to be minor and short-lived. The North-North Temperate Zone (NNTZ) is perhaps more prominent, but also generally quiet. Other minor belts and zones in the region are occasionally observed.^[34]

The North Temperate Region is part of a latitudinal region easily observable from Earth, and thus has a superb record of observation.^[35] It also features the strongest prograde jet stream on the planet—a westerly current that forms the southern boundary of the North Temperate Belt (NTB).^[35] The NTB fades roughly once a decade (this was the case during the *Voyager* encounters), making the North Temperate Zone (NTZ) apparently merge into the North Tropical Zone (NTropZ).^[35] Other times, the NTZ is divided by a narrow belt into northern and southern components.^[35]



Idealized illustration of Jupiter's cloud bands, labeled with their official abbreviations. Lighter zones are indicated to the right, darker belts to the left. The Great Red Spot and Oval BA are shown in the South Tropical Zone and South Temperate Belt, respectively.

The North Tropical Region is composed of the NTropZ and the North Equatorial Belt (NEB). The NTropZ is generally stable in coloration, changing in tint only in tandem with activity on the NTB's southern jet stream. Like the NTZ, it too is sometimes divided by a narrow band, the NTropB. On rare occasions, the southern NTropZ plays host to "Little Red Spots". As the name suggests, these are northern equivalents of the Great Red Spot. Unlike the GRS, they tend to occur in pairs and are always short-lived, lasting a year on average; one was present during the *Pioneer 10* encounter.^[36]

The NEB is one of the most active belts on the planet. It is characterized by anticyclonic white ovals and cyclonic "barges" (also known as "brown ovals"), with the former usually forming farther north than the latter; as in the NTropZ, most of these features are relatively short-lived. Like the South Equatorial Belt (SEB), the NEB has sometimes dramatically faded and "revived". The timescale of these changes is about 25 years.^[37]

The Equatorial Region (EZ) is one of the most stable regions of the planet, in latitude and in activity. The northern edge of the EZ hosts spectacular plumes that trail southwest from the NEB, which are bounded by dark, warm (in infrared) features known as festoons (hot spots).^[38] Though the southern boundary of the EZ is usually quiescent, observations from the late 19th into the early 20th century show that this pattern was then reversed relative to today. The EZ varies considerably in coloration, from pale to an



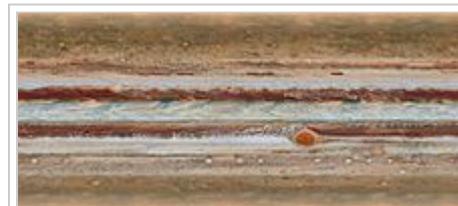
Zones, belts and vortices on Jupiter: The wide equatorial zone is visible in the center surrounded by two dark equatorial belts (SEB and NEB). The large grayish-blue irregular "hot spots" at the northern edge of the white Equatorial Zone change over the course of time as they march eastward across the planet. The Great Red Spot is at the southern margin of the SEB. Strings of small storms rotate around northern-hemisphere ovals. Small, very bright features, possible lightning storms, appear quickly and randomly in turbulent regions. The smallest features visible at the equator are about 600 kilometers across. This 14-frame animation spans 24 Jovian days, or about 10 Earth days. The passage of time is accelerated by a factor of 600,000.

ochre, or even coppery hue; it is occasionally divided by an Equatorial Band (EB).^[39] Features in the EZ move roughly 390 km/h relative to the other latitudes.^{[40][41]}

The South Tropical Region includes the South Equatorial Belt (SEB) and the South Tropical Zone. It is by far the most active region the planet, as it is home to its strongest retrograde jet stream. The SEB is usually the broadest, darkest belt on Jupiter; it is sometimes split by a zone (the SEBZ), and can fade entirely every 3 to 15

years before reappearing in what is known as an SEB Revival cycle. A period of weeks or months following the belt's disappearance, a white spot forms and erupts dark brownish material which is stretched into a new belt by Jupiter's winds. The belt most recently disappeared in May 2010.^[42] Another characteristic of the SEB is a long train of cyclonic disturbances following the Great Red Spot. Like the NTropZ, the STropZ is one of the most prominent zones on the planet; not only does it contain the GRS, but it is occasionally rent by a South Tropical Disturbance (STropD), a division of the zone that can be very long-lived; the most famous one lasted from 1901 to 1939.^[43]

The South Temperate Region, or South Temperate Belt (STB), is yet another dark, prominent belt, more so than the NTB; until March 2000, its most famous features were the long-lived white ovals BC, DE, and FA, which have since merged to form Oval BA ("Red Jr."). The ovals were part of South Temperate Zone, but they extended into STB partially blocking it.^[6] The STB has occasionally faded, apparently due to complex interactions between the white ovals and the GRS. The appearance of the South Temperate Zone (STZ)—the zone in which the white ovals originated—is highly variable.^[45]

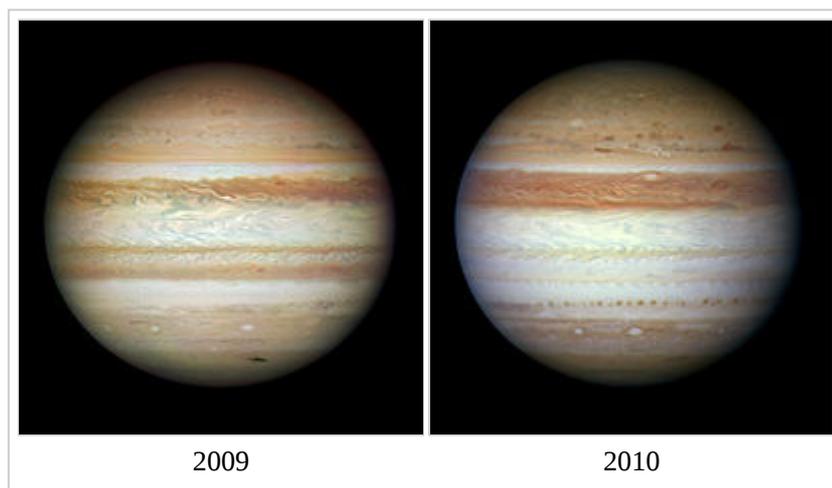


Images also reveal a rare wave structure in the planet's atmosphere.^[44]

There are other features on Jupiter that are either temporary or difficult to observe from Earth. The South South Temperate Region is harder to discern even than the NNTR; its detail is subtle and can only be studied well by large telescopes or spacecraft.^[46] Many zones and belts are more transient in nature and are not always visible. These include the *Equatorial band* (EB),^[47] *North Equatorial belt zone* (NEBZ, a white zone within the belt) and *South Equatorial belt zone* (SEBZ).^[48] Belts are also occasionally split by a sudden disturbance. When a disturbance divides a normally singular belt or zone, an *N* or an *S* is added to indicate whether the component is the northern or southern one; e.g., NEB(N) and NEB(S).^[49]

Dynamics

Circulation in Jupiter's atmosphere is markedly different from that in the atmosphere of Earth. The interior of Jupiter is fluid and lacks any solid surface. Therefore, convection may occur throughout the planet's outer molecular envelope. As of 2008, a comprehensive theory of the dynamics of the Jovian atmosphere has not been developed. Any such theory needs to explain the following facts: the existence of narrow stable bands and jets that are symmetric relative to Jupiter's equator, the strong prograde jet observed at the equator, the difference



between zones and belts, and the origin and persistence of large vortices such as the Great Red Spot.^[7]

The theories regarding the dynamics of the Jovian atmosphere can be broadly divided into two classes: shallow and deep. The former hold that the observed circulation is largely confined to a thin outer (weather) layer of the planet, which overlays the stable interior. The latter hypothesis postulates that the observed atmospheric flows are only a surface manifestation of deeply rooted circulation in the outer molecular envelope of Jupiter.^[50] As both theories have their own successes and failures, many planetary scientists think that the true theory will include elements of both models.^[51]

Shallow models

The first attempts to explain Jovian atmospheric dynamics date back to the 1960s.^{[50][52]} They were partly based on terrestrial meteorology, which had become well developed by that time. Those shallow models assumed that the jets on Jupiter are driven by small scale turbulence, which is in turn maintained by moist convection in the outer layer of the atmosphere (above the water clouds).^{[53][54]} The moist convection is a phenomenon related to the condensation and evaporation of water and is one of the major drivers of terrestrial weather.^[55] The production of the jets in this model is related to a well-known property of two dimensional turbulence—the so-called inverse cascade, in which small turbulent structures (vortices) merge to form larger ones.^[53] The finite size of the planet means that the cascade can not produce structures larger than some characteristic scale, which for Jupiter is called the Rhines scale. Its existence is connected to production of Rossby waves. This process works as follows: when the largest turbulent structures reach a certain size, the energy begins to flow into Rossby waves instead of larger structures, and the inverse cascade stops.^[56] Since on the spherical rapidly rotating planet the dispersion relation of the Rossby waves is anisotropic, the Rhines scale in the direction parallel to the equator is larger than in the direction orthogonal to it.^[56] The ultimate result of the process described above is production of large scale elongated structures, which are parallel to the equator. The meridional extent of them appears to match the actual width of jets.^[53] Therefore, in shallow models vortices actually feed the jets and should disappear by merging into them.

While these weather–layer models can successfully explain the existence of a dozen narrow jets, they have serious problems.^[53] A glaring failure of the model is the prograde (super-rotating) equatorial jet: with some rare exceptions shallow models produce a strong retrograde (subrotating) jet, contrary to observations. In addition, the jets tend to be unstable and can disappear over time.^[53] Shallow models cannot explain how the observed atmospheric flows on Jupiter violate stability criteria.^[57] More elaborated multilayer versions of weather–layer models produce more stable circulation, but many problems persist.^[58] Meanwhile, the Galileo Probe found that the winds on Jupiter extend well below the water clouds at 5–7 bar and do not show any evidence of decay down to 22 bar pressure level, which implies that circulation in the Jovian atmosphere may in fact be deep.^[17]

Deep models

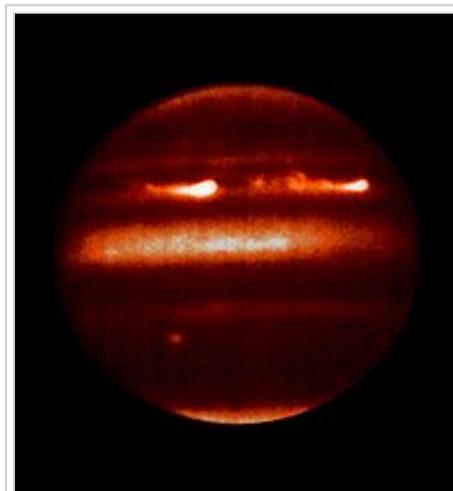
The deep model was first proposed by Busse in 1976.^{[59][60]} His model was based on another well-known feature of fluid mechanics, the Taylor–Proudman theorem. It holds that in any fast-rotating barotropic ideal liquid, the flows are organized in a series of cylinders parallel to the rotational axis. The conditions of the theorem are probably met in the fluid Jovian interior. Therefore, the planet's molecular hydrogen mantle may be divided into cylinders, each cylinder having a circulation independent of the others.^[61] Those latitudes where the cylinders' outer and inner boundaries intersect with the visible surface of the planet correspond to the jets; the cylinders themselves are observed as zones and belts.

The deep model easily explains the strong prograde jet observed at the equator of Jupiter; the jets it produces are stable and do not obey the 2D stability criterion.^[61] However it has major difficulties; it produces a very small number of broad jets, and realistic simulations of 3D flows are not possible as of 2008, meaning that the simplified models used to justify deep circulation may fail to catch important aspects of the fluid dynamics within Jupiter.^[61] One model published in 2004 successfully reproduced the Jovian band-jet structure.^[51] It assumed that the molecular hydrogen mantle is thinner than in all other models; occupying only the outer 10% of Jupiter's radius. In standard models of the Jovian interior, the mantle comprises the outer 20–30%.^[62] The driving of deep circulation is another problem. The deep flows can be caused both by shallow forces (moist convection, for instance) or by deep planet-wide convection that transports heat out of the Jovian interior.^[53] Which of these mechanisms is more important is not clear yet.

Internal heat

As has been known since 1966,^[63] Jupiter radiates much more heat than it receives from the Sun. It is estimated that the ratio between the power emitted by the planet and that absorbed from the Sun is 1.67 ± 0.09 . The internal heat flux from Jupiter is $5.44 \pm 0.43 \text{ W/m}^2$, whereas the total emitted power is 335 ± 26 petawatts. The latter value is approximately equal to one billionth of the total power radiated by the Sun. This excess heat is mainly the primordial heat from the early phases of Jupiter's formation, but may result in part from the precipitation of helium into the core.^[64]

The internal heat may be important for the dynamics of the Jovian atmosphere. While Jupiter has a small obliquity of about 3° , and its poles receive much less solar radiation than its equator, the tropospheric temperatures do not change appreciably from the equator to poles. One explanation is that Jupiter's convective interior acts like a thermostat, releasing more heat near the poles than in the equatorial region. This leads to a uniform temperature in the troposphere. While heat is transported from the equator to the poles mainly via the atmosphere on Earth, on Jupiter deep convection equilibrates heat. The convection in the Jovian interior is thought to be driven mainly by the internal heat.^[65]



Thermal image of Jupiter obtained by NASA Infrared Telescope Facility

Discrete features

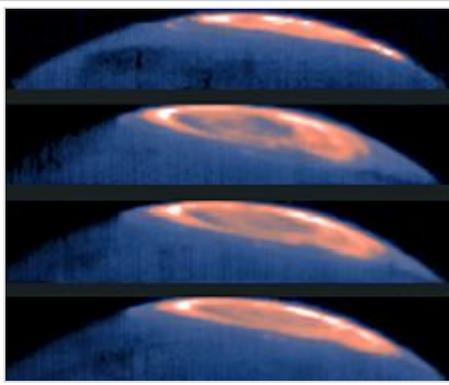
Vortices

The atmosphere of Jupiter is home to hundreds of vortices—circular rotating structures that, as in the Earth's atmosphere, can be divided into two classes: cyclones and anticyclones.^[8] Cyclones rotate in the direction similar to the rotation of the planet (counterclockwise in the northern hemisphere and clockwise in the southern); the anticyclones rotate in the reverse direction. However a major difference from the terrestrial atmosphere is that, in the Jovian atmosphere, anticyclones dominate over cyclones, as more than 90% of vortices larger than 2000 km in diameter are anticyclones.^[66] The lifetime of vortices varies from several days to hundreds of years depending on their size. For instance, the average lifetime of anticyclones with diameters from 1000 to 6000 km is 1–3 years.^[67] Vortices have never been observed in the equatorial region of Jupiter (within 10° of latitude), where they are unstable.^[11] As on any rapidly rotating planet, Jupiter's anticyclones are high pressure centers, while cyclones are low pressure.^[38]

The anticyclones in Jupiter's atmosphere are always confined within zones, where the wind speed increases in direction from the equator to the poles.^[67] They are usually bright and appear as white ovals.^[8] They can move in longitude, but stay at approximately the same latitude as they are unable to escape from the confining zone.^[11] The wind speeds at their periphery are about 100 m/s.^[10] Different anticyclones located in one zone tend to merge, when they approach each other.^[69] However Jupiter has two anticyclones that are somewhat different from all others. They are the Great Red Spot (GRS)^[9] and the Oval BA,^[10] the latter formed only in 2000. In contrast to white ovals, these structures are red, arguably due to dredging up of red material from the planet's depths.^[9] On Jupiter the anticyclones usually form through merges of smaller



New Horizons IR view of Jupiter's atmosphere



Great Cold Spot on Jupiter.^[68]

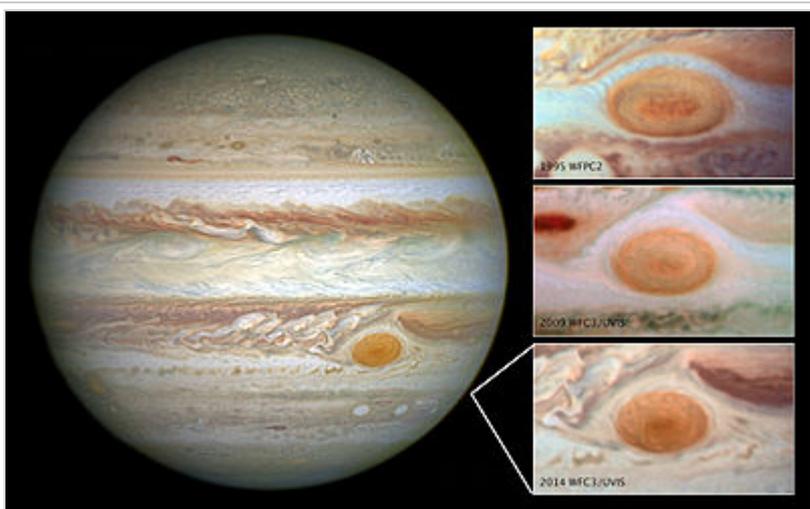
structures including convective storms (see below),^[67] although large ovals can result from the instability of jets. The latter was observed in 1938–1940, when a few white ovals appeared as a result of instability of the southern temperate zone; they later merged to form Oval BA.^{[10][67]}

In contrast to anticyclones, the Jovian cyclones tend to be small, dark and irregular structures. Some of the darker and more regular features are known as brown ovals (or badges).^[66] However the existence of a few long-lived large cyclones has been suggested. In addition to compact cyclones, Jupiter has several large irregular filamentary patches, which demonstrate cyclonic rotation.^[8] One of them is located to the west of the GRS (in its wake region) in the southern equatorial belt.^[70] These patches are called cyclonic regions (CR). The cyclones

are always located in the belts and tend to merge when they encounter each other, much like anticyclones.^[67]

The deep structure of vortices is not completely clear. They are thought to be relatively thin, as any thickness greater than about 500 km will lead to instability. The large anticyclones are known to extend only a few tens of kilometers above the visible clouds. The early hypothesis that the vortices are deep convective plumes (or convective columns) as of 2008 is not shared by the majority of planetary scientists.^[11]

Great Red Spot



Jupiter – Great Red Spot is decreasing in size (May 15, 2014).^[71]

The Great Red Spot (GRS) is a persistent anticyclonic storm, 22° south of Jupiter's equator; observations from Earth establish a minimum storm lifetime of 350 years.^{[72][73]} A storm was described as a "permanent spot" by Gian Domenico Cassini after observing the feature in July 1665 with his instrument-maker Eustachio Divini.^[74] According to a report by Giovanni Battista Riccioli in 1635, Leander Bandtius, whom Riccioli identified as the Abbot of Dunisburgh who possessed an "extraordinary telescope", observed a large spot that he described as "oval, equaling one seventh of Jupiter's diameter at its longest." According to Riccioli, "these

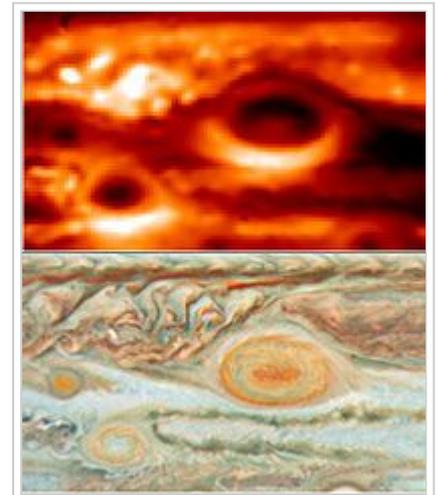
features are seldom able to be seen, and then only by a telescope of exceptional quality and magnification."^[75] The Great Spot has been nearly continually observed since the 1870s, however.

The GRS rotates counter-clockwise, with a period of about six Earth days^[76] or 14 Jovian days. Its dimensions are 24,000–40,000 km east-to-west and 12,000–14,000 km north-to-south. The spot is large enough to contain two or three planets the size of Earth. At the start of 2004, the Great Red Spot had approximately half the longitudinal extent it had a century ago, when it was 40,000 km in diameter. At the present rate of reduction, it could potentially become circular by 2040, although this is unlikely because of the distortion effect of the neighboring jet streams.^[77] It is not known how long the spot will last, or whether the change is a result of normal fluctuations.^[78]

According to a study by scientists at the University of California, Berkeley, between 1996 and 2006 the spot lost 15 percent of its diameter along its major axis. Xylar Asay-Davis, who was on the team that conducted the study, noted that the spot is not disappearing because "velocity is a more robust measurement because the

clouds associated with the Red Spot are also strongly influenced by numerous other phenomena in the surrounding atmosphere."^[79]

Infrared data have long indicated that the Great Red Spot is colder (and thus, higher in altitude) than most of the other clouds on the planet;^[80] the cloudtops of the GRS are about 8 km above the surrounding clouds. Furthermore, careful tracking of atmospheric features revealed the spot's counterclockwise circulation as far back as 1966 – observations dramatically confirmed by the first time-lapse movies from the *Voyager* flybys.^[81] The spot is spatially confined by a modest eastward jet stream (prograde) to its south and a very strong westward (retrograde) one to its north.^[82] Though winds around the edge of the spot peak at about 120 m/s (432 km/h), currents inside it seem stagnant, with little inflow or outflow.^[83] The rotation period of the spot has decreased with time, perhaps as a direct result of its steady reduction in size.^[84] In 2010, astronomers imaged the GRS in the far infrared (from 8.5 to 24 μm) with a spatial resolution higher than ever before and found that its central, reddest region is warmer than its surroundings by between 3–4 K. The warm airmass is located in the upper troposphere in the pressure range of 200–500 mbar. This warm central spot slowly counter-rotates and may be caused by a weak subsidence of air in the center of GRS.^[85]



An infrared image of GRS (top) and Oval BA (lower left) showing its cool center, taken by the ground based Very Large Telescope. An image made by the Hubble Space Telescope (bottom) is shown for comparison.

The Great Red Spot's latitude has been stable for the duration of good observational records, typically varying by about a degree. Its longitude, however, is subject to constant variation.^{[86][87]} Because Jupiter's visible features do not rotate uniformly at all latitudes, astronomers have defined three different systems for defining the longitude. System II is used for latitudes of more than 10°, and was originally based on the average rotation rate of the Great Red Spot of 9h 55m 42s.^{[88][89]} Despite this, the spot has 'lapped' the planet in System II at least 10 times since the early 19th century. Its drift rate has changed dramatically over the years and has been linked to the brightness of the South Equatorial Belt, and the presence or absence of a South Tropical Disturbance.^[90]



Approximate size comparison of Earth superimposed on this Dec 29, 2000 image showing the Great Red Spot

It is not known exactly what causes the Great Red Spot's reddish color. Theories supported by laboratory experiments suppose that the color may be caused by complex organic molecules, red phosphorus, or yet another sulfur compound. The GRS varies greatly in hue, from almost brick-red to pale salmon, or even white. The higher temperature of the reddest central region is the first evidence that the Spot's color is affected by environmental factors.^[85] The spot occasionally disappears from the visible spectrum, becoming evident only through the Red Spot Hollow, which is its niche in the South Equatorial Belt (SEB). The visibility of GRS is apparently coupled to the appearance of the SEB; when the belt is bright white, the spot tends to be dark, and when it is dark, the spot is usually light. The periods when the spot is dark or light occur at irregular intervals; in the 50 years from 1947 to 1997, the spot was darkest in the periods 1961–1966, 1968–1975, 1989–1990, and 1992–1993.^[91] In November 2014, an analysis of data

from NASA's Cassini mission revealed that the red color is likely a product of simple chemicals being broken apart by sunlight in the planet's upper atmosphere ^{[92][93]}

The Great Red Spot should not be confused with the Great Dark Spot, a feature observed near Jupiter's north pole in 2000 by the Cassini–Huygens spacecraft.^[94] A feature in the atmosphere of Neptune was also called the Great Dark Spot. The latter feature, imaged by *Voyager 2* in 1989, may have been an atmospheric hole rather than a storm. It was no longer present in 1994, although a similar spot had appeared farther to the north.^[95]

Oval BA

Oval BA is a red storm in Jupiter's southern hemisphere similar in form to, though smaller than, the Great Red Spot (it is often affectionately referred to as "Red Spot Jr.", "Red Jr." or "The Little Red Spot"). A feature in the South Temperate Belt, Oval BA was first seen in 2000 after the collision of three small white storms, and has intensified since then.^[96]

The formation of the three white oval storms that later merged into Oval BA can be traced to 1939, when the South Temperate Zone was torn by dark features that effectively split the zone into three long sections.

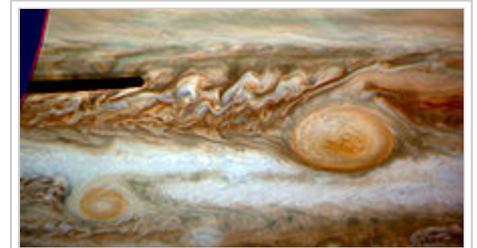
Jovian observer Elmer J. Reese labeled the dark sections AB, CD, and EF. The rifts expanded, shrinking the remaining segments of the STZ into the white ovals FA, BC, and DE.^[97] Ovals BC and DE merged in 1998, forming Oval BE. Then, in March 2000, BE and FA joined together, forming Oval BA.^[96] (see White ovals, below)

Oval BA slowly began to turn red in August 2005.^[98] On February 24, 2006, Filipino amateur astronomer Christopher Go discovered the color change, noting that it had reached the same shade as the GRS.^[98] As a result, NASA writer Dr. Tony Phillips suggested it be called "Red Spot Jr." or "Red Jr."^[99]

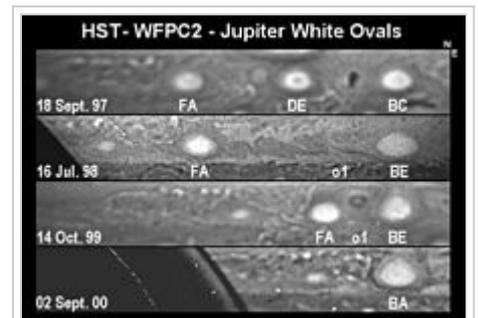
In April 2006, a team of astronomers, believing that Oval BA might converge with the GRS that year, observed the storms through the Hubble Space Telescope.^[100] The storms pass each other about every two years, but the passings of 2002 and 2004 did not produce anything exciting. Dr. Amy Simon-Miller, of the Goddard Space Flight Center, predicted the storms would have their closest passing on July 4, 2006.^[100] On July 20, the two storms were photographed passing each other by the Gemini Observatory without converging.^[101]

Why Oval BA turned red is not understood. According to a 2008 study by Dr. Santiago Pérez-Hoyos of the University of the Basque Country, the most likely mechanism is "an upward and inward diffusion of either a colored compound or a coating vapor that may interact later with high energy solar photons at the upper levels of Oval BA."^[102] Some believe that small storms (and their corresponding white spots) on Jupiter turn red when the winds become powerful enough to draw certain gases from deeper within the atmosphere which change color when those gases are exposed to sunlight.^[103]

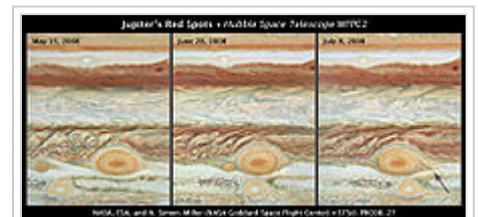
Oval BA is getting stronger according to observations made with the Hubble Space Telescope in 2007. The wind speeds have reached 618 km/h; about the same as in the Great Red Spot and far stronger than any of the progenitor storms.^{[104][105]} As of July 2008, its size is about the diameter of Earth—approximately half the size of the Great Red Spot.^[102]



Oval BA (left)



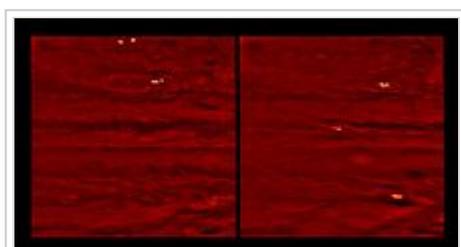
Formation of Oval BA from three white ovals



Oval BA (bottom), Great Red Spot (top) and "Baby Red Spot" (middle) during a brief encounter in June, 2008

Oval BA should not be confused with another major storm on Jupiter, the South Tropical Little Red Spot (LRS) (nicknamed "the Baby Red Spot" by NASA^[106]), which was destroyed by the GRS.^[103] The new storm, previously a white spot in Hubble images, turned red in May 2008. The observations were led by Imke de Pater of the University of California, at Berkeley, US.^[107] The Baby Red Spot encountered the GRS in late June to early July 2008, and in the course of a collision, the smaller red spot was shredded into pieces. The remnants of the Baby Red Spot first orbited, then were later consumed by the GRS. The last of the remnants with a reddish color to have been identified by astronomers had disappeared by mid-July, and the remaining pieces again collided with the GRS, then finally merged with the bigger storm. The remaining pieces of the Baby Red Spot had completely disappeared by August 2008.^[106] During this encounter Oval BA was present nearby, but played no apparent role in destruction of the Baby Red Spot.^[106]

Storms and lightning



Lightning on Jupiter's night side, imaged by the Galileo orbiter in 1997

The storms on Jupiter are similar to thunderstorms on Earth. They reveal themselves via bright clumpy clouds about 1000 km in size, which appear from time to time in the belts' cyclonic regions, especially within the strong westward (retrograde) jets.^[109] In contrast to vortices, storms are short-lived

phenomena; the strongest of them may exist for several months, while the average lifetime is only 3–4 days.^[109] They are believed to be due mainly to moist convection within Jupiter's troposphere. Storms are actually tall convective columns (plumes), which bring the wet air from the depths to the upper part of the troposphere, where it condenses in clouds. A typical vertical extent of Jovian storms is about 100 km; as they extend from a pressure level of about 5–7 bar, where the base of a hypothetical water cloud layer is located, to as high as 0.2–0.5 bar.^[110]

Storms on Jupiter are always associated with lightning. The imaging of the night-side hemisphere of Jupiter by *Galileo* and *Cassini* spacecraft revealed regular light flashes in Jovian belts and near the locations of the westward jets, particularly at 51°N, 56°S and 14°S latitudes.^[111] On Jupiter lightning strikes are on average a few times more powerful than those on Earth. However, they are less frequent; the light power emitted from a given area is similar to that on Earth.^[111] A few flashes have been detected in polar regions, making Jupiter the second known planet after Earth to exhibit polar lightning.^[112]

Every 15–17 years Jupiter is marked by especially powerful storms. They appear at 23°N latitude, where the strongest eastward jet, that can reach 150 m/s, is located. The last time such an event was observed was in March–June 2007.^[110] Two storms appeared in the northern temperate belt 55° apart in longitude. They significantly disturbed the belt. The dark material that was shed by the storms mixed with clouds and changed the belt's color. The storms moved with a speed as high as 170 m/s, slightly faster than the jet itself, hinting at the existence of strong winds deep in the atmosphere.^[110]

Disturbances

The normal pattern of bands and zones is sometimes disrupted for periods of time. One particular class of disruption are long-lived darkenings of the South Tropical Zone, normally referred to as "South Tropical Disturbances" (STD). The longest lived STD in recorded history was followed from 1901 until 1939, having



Jupiter – southern storms – JunoCam.^[108]

been first seen by Percy B. Molesworth on February 28, 1901. It took the form of darkening over part of the normally bright South Tropical zone. Several similar disturbances in the South Tropical Zone have been recorded since then.^[113]

Hot spots

One of the most mysterious features in the atmosphere of Jupiter are hot spots. In them the air is relatively free of clouds and heat can escape from the depths without much absorption. The spots look like bright spots in the infrared images obtained at the wavelength of about $5\ \mu\text{m}$.^[38] They are preferentially located in the belts, although there is a train of prominent hot spots at the northern edge of the Equatorial Zone. The Galileo Probe descended into one of those equatorial spots. Each equatorial spot is associated with a bright cloudy plume located to the west of it and reaching up to 10,000 km in size.^[6] Hot spots generally have round shapes, although they do not resemble vortices.^[38]

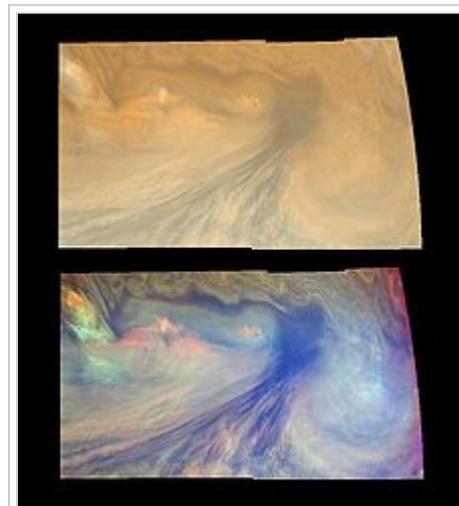
The origin of hot spots is not clear. They can be either downdrafts, where the descending air is adiabatically heated and dried or, alternatively, they can be a manifestation of planetary scale waves. The latter hypothesis explains the periodical pattern of the equatorial spots.^{[6][38]}

Observational history

Early astronomers, using small telescopes, recorded the changing appearance of Jupiter's atmosphere.^[23] Their descriptive terms—belts and zones, brown spots and red spots, plumes, barges, festoons, and streamers—are still used.^[114] Other terms such as vorticity, vertical motion, cloud heights have entered in use later, in the 20th century.^[23]

The first observations of the Jovian atmosphere at higher resolution than possible with Earth-based telescopes were taken by the *Pioneer 10* and *11* spacecraft. The first truly detailed images of Jupiter's atmosphere were provided by the *Voyagers*.^[23] The two spacecraft were able to image details at a resolution as low as 5 km in size in various spectra, and also able to create "approach movies" of the atmosphere in motion.^[23] The Galileo Probe, which suffered an antenna problem, saw less of Jupiter's atmosphere but at a better average resolution and a wider spectral bandwidth.^[23]

Today, astronomers have access to a continuous record of Jupiter's atmospheric activity thanks to telescopes such as Hubble Space Telescope. These show that the atmosphere is occasionally wracked by massive disturbances, but that, overall, it is remarkably stable.^[23] The vertical motion of Jupiter's atmosphere was largely determined by the identification of trace gases by ground-based telescopes.^[23] Spectroscopic studies after the collision of Comet Shoemaker–Levy 9 gave a glimpse of the Jupiter's composition beneath the cloud tops. The presence of diatomic sulfur (S_2) and carbon disulfide (CS_2) was recorded—the first detection of either in Jupiter, and only the second detection of S_2 in any astronomical object— together with other molecules such as ammonia (NH_3) and hydrogen sulfide (H_2S), while oxygen-bearing molecules such as sulfur dioxide were not detected, to the surprise of astronomers.^[115]



False color image of an equatorial hot spot



Time-lapse sequence from the approach of *Voyager 1* to Jupiter

The *Galileo* atmospheric probe, as it plunged into Jupiter, measured the wind, temperature, composition, clouds, and radiation levels down to 22 bar. However, below 1 bar elsewhere on Jupiter there is uncertainty in the quantities.^[23]

Great Red Spot studies

The first sighting of the GRS is often credited to Robert Hooke, who described a spot on the planet in May 1664; however, it is likely that Hooke's spot was in the wrong belt altogether (the North Equatorial Belt, versus the current location in the South Equatorial Belt). Much more convincing is Giovanni Cassini's description of a "permanent spot" in the following year.^[116] With fluctuations in visibility, Cassini's spot was observed from 1665 to 1713.^[117]

A minor mystery concerns a Jovian spot depicted around 1700 on a canvas by Donato Creti, which is exhibited in the Vatican.^{[118][119]} It is a part of a series of panels in which different (magnified) heavenly bodies serve as backdrops for various Italian scenes, the creation of all of them overseen by the astronomer Eustachio Manfredi for accuracy. Creti's painting is the first known to depict the GRS as red. No Jovian feature was officially described as red before the late 19th century.^[119]

The present GRS was first seen only after 1830 and well-studied only after a prominent apparition in 1879. A 118-year gap separates the observations made after 1830 from its 17th-century discovery; whether the original spot dissipated and re-formed, whether it faded, or even if the observational record was simply poor are unknown.^[91] The older spots had a short observational history and slower motion than that of the modern spot, which make their identity unlikely.^[118]



A wider view of Jupiter and the Great Red Spot as seen from *Voyager 1* in 1979



Hubble's Wide Field Camera 3 took the GRS region at its smallest size ever.

On February 25, 1979, when the *Voyager 1* spacecraft was 9.2 million kilometers from Jupiter it transmitted the first detailed image of the Great Red Spot back to Earth. Cloud details as small as 160 km across were visible. The colorful, wavy cloud pattern seen to the west (left) of the GRS is the spot's wake region, where extraordinarily complex and variable cloud motions are observed.^[120]

White ovals

The white ovals that were to become Oval BA formed in 1939. They covered almost 90 degrees of longitude shortly after their formation, but contracted rapidly during their first decade; their length stabilized at 10 degrees or less

after 1965.^[121] Although they originated as segments of the STZ, they evolved to become completely embedded in the South Temperate Belt, suggesting that they moved north, "digging" a niche into the STB.^[122] Indeed, much like the GRS, their circulations were confined by two opposing jet streams on their northern and southern boundaries, with an eastward jet to their north and a retrograde westward one to the south.^[121]



The white ovals that later formed Oval BA, imaged by the *Galileo* orbiter in 1997

The longitudinal movement of the ovals seemed to be influenced by two factors: Jupiter's position in its orbit (they became faster at aphelion), and their proximity to the GRS (they accelerated when within 50 degrees of the Spot).^[123] The overall trend of the white oval drift rate was deceleration, with a decrease by half between 1940 and 1990.^[124]

During the *Voyager* fly-bys, the ovals extended roughly 9000 km from east to west, 5000 km from north to south, and rotated every five days (compared to six for the GRS at the time).^[125]

See also

- Comet Shoemaker–Levy 9
- Extrasolar planet (many larger than Jupiter)
- Galileo Spacecraft (a mission that included both an orbiter and an atmospheric-entry probe)
- Juno probe
- 2009 Jupiter impact event
- 2010 Jupiter impact event
- Ulysses (spacecraft)
- Voyager 1, Voyager 2



Wikimedia Commons has media related to ***Great Red Spot***.

Notes

- a. [^] The scale height sh is defined as $sh = RT/(Mg_j)$, where $R = 8.31 \text{ J/mol/K}$ is the gas constant, $M \approx 0.0023 \text{ kg/mol}$ is the average molar mass in the Jovian atmosphere,^[4] T is temperature and $g_j \approx 25 \text{ m/s}^2$ is the gravitational acceleration at the surface of Jupiter. As the temperature varies from 10 K in the tropopause up to 1000 K in the thermosphere,^[4] the scale height can assume values from 15 to 150 km.
- b. [^] The *Galileo* atmospheric probe failed to measure the deep abundance of oxygen, because the water concentration continued to increase down to the pressure level of 22 bar when it ceased operating. While the actually measured oxygen abundances are much lower than the solar value, the observed rapid increase of water content of the atmosphere with depth makes it highly likely that the deep abundance of oxygen indeed exceeds the solar value by a factor of about 3—much like other elements.^[2]
- c. [^] Various explanations of the overabundance of carbon, oxygen, nitrogen and other elements have been proposed. The leading one is that Jupiter captured a large number of icy planetesimals during the later stages of its accretion. The volatiles like noble gases are thought to have been trapped as clathrate hydrates in water ice.^[2]

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External links

-  Media related to Atmosphere of Jupiter at Wikimedia Commons

- Planetary Society blog post (2017-05-09) by Peter Rosén describing assembly of a video of Jupiter's atmospheric activity from 19 December 2014 to 31 March 2015 from amateur astronomer images

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Atmosphere of Mars

From Wikipedia, the free encyclopedia

The **atmosphere of Mars** is the layer of gases surrounding Mars. It is composed mostly of carbon dioxide. The atmospheric pressure on the Martian surface averages 600 pascals (0.087 psi; 6.0 mbar), about 0.6% of Earth's mean sea level pressure of 101.3 kilopascals (14.69 psi; 1.013 bar). It ranges from a low of 30 pascals (0.0044 psi; 0.30 mbar) on Olympus Mons's peak to over 1,155 pascals (0.1675 psi; 11.55 mbar) in the depths of Hellas Planitia. This pressure is well below the Armstrong limit for the unprotected human body. Mars's atmospheric mass of 25 teratonnes compares to Earth's 5148 teratonnes with a scale height of about 11 kilometres (6.8 mi) versus Earth's 7 kilometres (4.3 mi).

The Martian atmosphere consists of approximately 96% carbon dioxide, 1.9% argon, 1.9% nitrogen, and traces of free oxygen, carbon monoxide, water and methane, among other gases,^[1] for a mean molar mass of 43.34 g/mol.^{[2][3]} There has been renewed interest in its composition since the detection of traces of methane in 2003^{[4][5]} that may indicate life but may also be produced by a geochemical process, volcanic or hydrothermal activity.^[6]

The atmosphere is quite dusty, giving the Martian sky a light brown or orange-red color when seen from the surface; data from the Mars Exploration Rovers indicate suspended particles of roughly 1.5 micrometres in diameter.^[7]

On 16 December 2014, NASA reported detecting an unusual increase, then decrease, in the amounts of methane in the atmosphere of the planet Mars. Organic chemicals have been detected in powder drilled from a rock by the *Curiosity* rover. Based on deuterium to hydrogen ratio studies, much of the water at Gale Crater on Mars was found to have been lost during ancient times, before the lakebed in the crater was formed; afterwards, large amounts of water continued to be lost.^{[8][9][10]}

On 18 March 2015, NASA reported the detection of an aurora that is not fully understood and an unexplained dust cloud in the atmosphere of Mars.^[11]

On 4 April 2015, NASA reported studies, based on measurements by the Sample Analysis at Mars (SAM) instrument on the *Curiosity* rover, of the Martian atmosphere using xenon and argon isotopes. Results provided support for a "vigorous" loss of atmosphere early in the history of Mars and were consistent with an atmospheric signature found in bits of atmosphere captured in some Martian meteorites found on Earth.^[12] This was further supported by results from the MAVEN orbiter circling Mars, that the solar wind is responsible for stripping away the atmosphere of Mars over the years.^{[13][14][15]}

Atmosphere of Mars

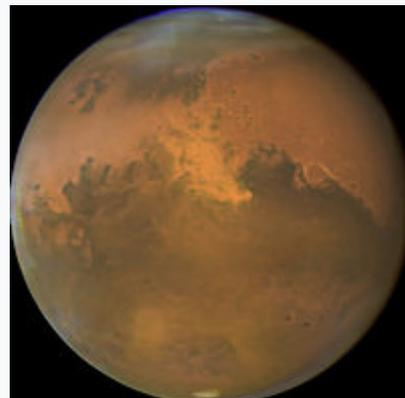


Image of Mars with sandstorm visible, taken by the *Hubble Space Telescope* on 28 October 2005

General information^[1]

Chemical species	Mole fraction
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Composition^[1]

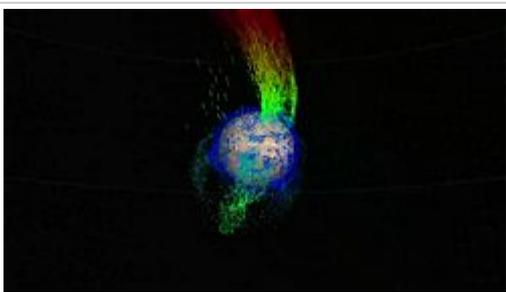
Carbon dioxide	95.97%
Argon	1.93%
Nitrogen	1.89%
Oxygen	0.146%
Carbon monoxide	0.0557%

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Structure



Play media

The solar wind accelerates ions from the Mars upper atmosphere into space (video (01:13); 5 November 2015)

Mars's atmosphere is composed of the following layers:

- **Exosphere:** Typically stated to start at 200 km (120 mi) and higher, this region is where the last wisps of atmosphere merge into the vacuum of space. There is no distinct boundary where the atmosphere ends; it just tapers away.
- **Upper atmosphere, or thermosphere:** A region with very high temperatures, caused by heating from the Sun. Atmospheric gases start to separate from each other at these altitudes, rather than forming the even mix found in the lower atmospheric layers.
- **Middle atmosphere:** The region in which Mars's jetstream flows.
- **Lower atmosphere:** A relatively warm region affected by heat from airborne dust and from the ground.

Pressure comparison

Where	Pressure
Olympus Mons summit	0.03 kilopascals (0.0044 psi)
Mars average	0.6 kilopascals (0.087 psi)
Hellas Planitia bottom	1.16 kilopascals (0.168 psi)
Armstrong limit	6.25 kilopascals (0.906 psi)
Mount Everest summit ^[16]	33.7 kilopascals (4.89 psi)
Earth sea level	101.3 kilopascals (14.69 psi)

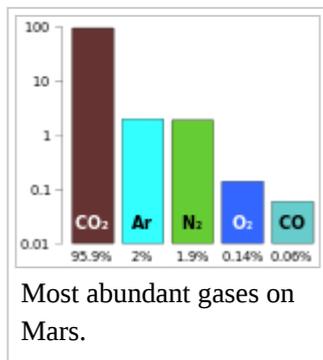
There is also a complicated ionosphere,^[17] and a seasonal ozone layer over the south pole.^[18] The *MAVEN* spacecraft determined in 2015 that there is a substantial layered structure present in both neutral gases and ion densities.^[19]

Observations and measurement from Earth

In 1864, William Rutter Dawes observed "that the ruddy tint of the planet does not arise from any peculiarity of its atmosphere seems to be fully proved by the fact that the redness is always deepest near the centre, where the atmosphere is thinnest."^[20] Spectroscopic observations in the 1860s and 1870s^{[21][22]} led many to think the atmosphere of Mars is similar to Earth's. In 1894, though, spectral analysis and other qualitative observations by William Wallace Campbell suggested Mars resembles the Moon, which has no appreciable atmosphere, in many respects.^[21]

In 1926, photographic observations by William Hammond Wright at the Lick Observatory allowed Donald Howard Menzel to discover quantitative evidence of Mars's atmosphere.^{[23][24]}

Composition



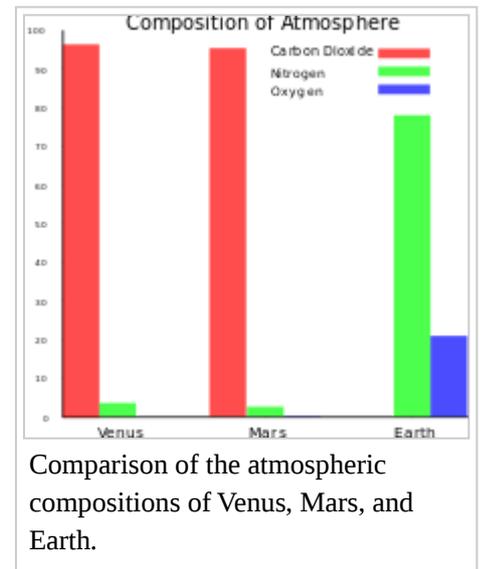
Carbon dioxide

The main component of the atmosphere of Mars is carbon dioxide (CO₂) at 95.9%. Each pole is in continual darkness during its hemisphere's winter, and the surface gets so cold that as much as 25% of the atmospheric CO₂ condenses at the polar caps into solid CO₂ ice (dry ice). When the pole is again exposed to sunlight during summer, the CO₂ ice sublimates back into the atmosphere. This process leads to a significant annual variation in the atmospheric pressure and atmospheric composition around the Martian poles.

It has been suggested that Mars had a much thicker, warmer, and wetter atmosphere early in its history.^[25] Much of this early atmosphere would have consisted of carbon dioxide. Such an atmosphere would have raised the temperature, at least in some places, to above the freezing point of water.^[26] With the higher temperature, running water could have carved out the many channels and outflow valleys that are common on the planet. It also might have gathered to form lakes and maybe an ocean.^[27] Some researchers have suggested that the atmosphere of Mars may have been many times as thick as the present one of Earth; however, research published in fall 2015 advanced the idea that perhaps the early Martian atmosphere was not as thick as previously thought.^[28] Currently, the atmosphere is very thin. For many years, it was assumed that as with Earth, most of the early carbon dioxide was locked up in minerals, called carbonates. However, despite the use of many orbiting instruments that looked for carbonates, very few carbonate deposits have been found.^{[28][29]} Today, it is thought that much of the carbon dioxide in the Martian air was removed by the solar wind. Researchers have discovered a two-step process that sends the gas into space.^[30] Ultraviolet light from the sun could strike a carbon dioxide molecule, breaking it into carbon monoxide and oxygen. A second photon of ultraviolet light could subsequently break the carbon monoxide into oxygen and carbon which would receive enough energy to escape the planet. In this process the light isotope of carbon (¹²C) is most likely to leave the atmosphere. Hence, the carbon dioxide left in the atmosphere would be enriched with the heavy isotope (¹³C).^[31] This higher level of the heavy isotope is what was recently found by the Curiosity Rover that sits on the surface of Mars.^{[32][33]}

Argon

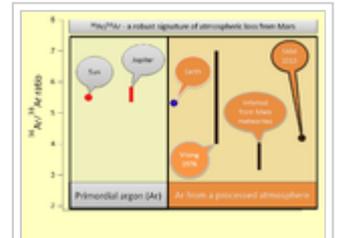
The atmosphere of Mars is enriched considerably with the noble gas argon, in comparison to the atmosphere of the other planets within the Solar System. Unlike carbon dioxide, the argon content of the atmosphere does not condense, and hence the total amount of argon in the Mars atmosphere is constant. However, the relative



concentration at any given location can change as carbon dioxide moves in and out of the atmosphere. Recent satellite data shows an increase in atmospheric argon over the southern pole during its autumn, which dissipates the following spring.^[36]

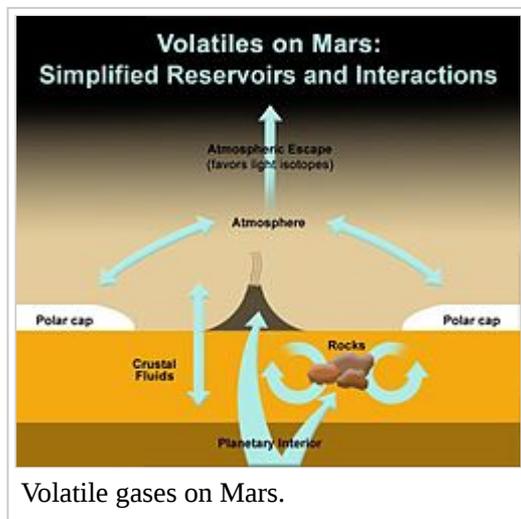
Water

Some aspects of the Martian atmosphere vary significantly. As carbon dioxide sublimates back into the atmosphere during the Martian summer, it leaves traces of water. Seasonal winds sweep off the poles at speeds approaching 400 kilometres per hour (250 mph) and transport large amounts of dust and water vapor giving rise to Earth-like frost and large cirrus clouds. These clouds of water-ice were photographed by the *Opportunity* rover in 2004.^[37] NASA scientists working on the *Phoenix* Mars mission confirmed on July 31, 2008 that they had indeed found subsurface water ice at Mars's northern polar region.



Argon isotope ratios are a signature of atmospheric loss on Mars.^{[34][35]}

Methane



Trace amounts of methane (CH₄), at the level of several parts per billion (ppb), were first reported in Mars's atmosphere by a team at the NASA Goddard Space Flight Center in 2003.^{[5][38]} In March 2004, the Mars Express Orbiter and ground-based observations by three groups also suggested the presence of methane in the atmosphere at a concentration of about 10 ppb (parts per billion).^{[39][40][41]} Large differences in the abundances were measured between observations taken in 2003 and 2006, which suggested that the methane was locally concentrated and probably seasonal.

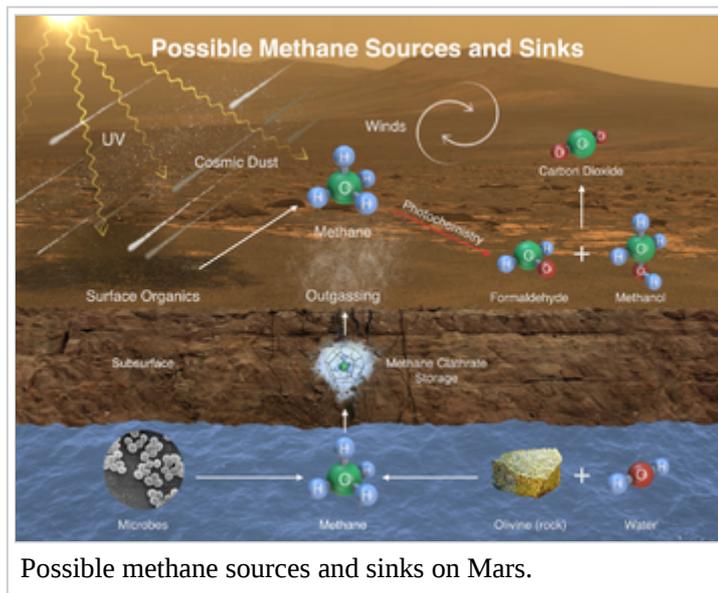
Because methane on Mars would quickly break down due to ultraviolet radiation from the Sun and chemical reactions with other gases, its reported persistent presence in the atmosphere also implies the existence of a source to continually replenish the gas.

Current photochemical models alone can not explain the rapid variability of the methane levels.^{[42][43]} It had been proposed that the methane might be replenished by meteorites entering the atmosphere of Mars,^[44] but researchers from Imperial College London found that the volumes of methane released this way are too low to sustain the measured levels of the gas.^[45]

Research suggests that the implied methane destruction lifetime is as long as ~4 Earth years and as short as ~0.6 Earth years.^{[46][47]} This lifetime is short enough for the atmospheric circulation to yield the observed uneven distribution of methane across the planet. In either case, the destruction lifetime for methane is much shorter than the timescale (~350 years) estimated for photochemical (UV radiation) destruction.^[46] The rapid destruction (or "sink") of methane suggests that another process must dominate removal of atmospheric methane on Mars, and it must be more efficient than destruction by light by a factor of 100 to 600.^{[47][46]} This unexplained fast destruction rate also suggests a very active replenishing source.^[48] In 2014 it was concluded that presence of strong methane sinks are not subject to atmospheric oxidation.^[49] A possibility is that the methane is not consumed at all, but rather condenses and evaporates seasonally from clathrates.^[50] Another possibility is that methane reacts with tumbling surface sand quartz (SiO₂) and olivine to form covalent Si-CH₃ bonds.^[51]

The principal candidates for the origin of Mars' methane include non-biological processes such as water-rock reactions, radiolysis of water, and pyrite formation, all of which produce H₂ that could then generate methane and other hydrocarbons via Fischer-Tropsch synthesis with CO and CO₂.^[52] It has also been shown that

methane could be produced by a process involving water, carbon dioxide, and the mineral olivine, which is known to be common on Mars.^[53] The required conditions for this reaction (i.e. high temperature and pressure) do not exist on the surface, but may exist within the crust.^{[54][55]} A detection of the mineral by-product serpentinite would suggest that this process is occurring. An analog on Earth suggests that low-temperature production and exhalation of methane from serpentinized rocks may be possible on Mars.^[56] Another possible geophysical source could be clathrate hydrates.^[57] Under the assumption of a cold early Mars environment, a cryosphere could trap such methane as clathrates in stable form at depth, that might exhibit sporadic release.^[58]



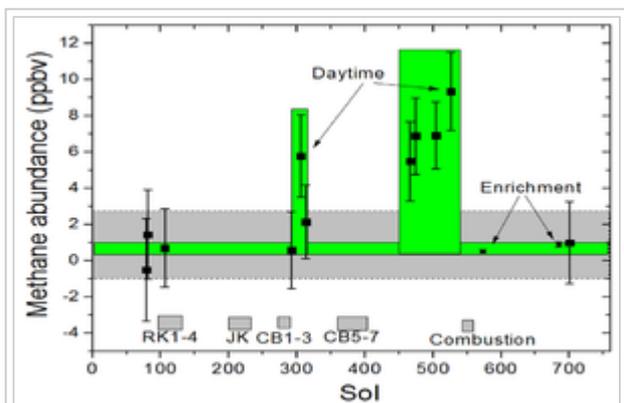
Possible methane sources and sinks on Mars.

A group of Mexican scientists performed plasma experiments in a synthetic Mars atmosphere and found that bursts of methane can be produced when a discharge interacts with water ice. A potential source of the discharges can be the electrification of dust particles from sand storms and dust devils. The ice can be found in trenches or in the permafrost. The electrical discharge ionizes gaseous CO₂ and water molecules and their byproducts recombine to produce methane. The results obtained show that pulsed electrical discharges over ice samples in a Martian atmosphere produce about 1.41×10^{16} molecules of methane per joule of applied energy.^{[59][60]}

Living microorganisms, such as methanogens, are another possible source, but no evidence exists for the presence of such organisms anywhere on Mars. In Earth's oceans, biological methane production tends to be accompanied by ethane, whereas volcanic methane is accompanied by sulfur dioxide.^[61] Several studies of trace gases in the Martian atmosphere have found no evidence for sulfur dioxide in the Martian atmosphere, which makes volcanism unlikely to be the source of methane.^{[62][63]}

In 2011, NASA scientists reported a comprehensive search using ground-based high-resolution infrared spectroscopy for trace species (including methane) on Mars, deriving sensitive upper limits for methane (<7 ppbv), ethane (<0.2 ppbv), methanol (<19 ppbv) and others (H₂CO, C₂H₂, C₂H₄, N₂O, NH₃, HCN, CH₃Cl, HCl, HO₂ – all limits at ppbv levels).^[64] The data were acquired over a period of 6 years and span different seasons and locations on Mars, suggesting that if organics are being released into the atmosphere, these events were extremely rare or currently non-existent, considering the expected long lifetimes for some of these species.^[64]

In August 2012, the Curiosity rover landed on Mars. The rover's instruments are capable of making precise abundance measurements, which can be used to distinguish between different isotopologues of methane.^{[65][66]} The first measurements with Curiosity's Tunable Laser Spectrometer (TLS) in 2012 indicated that there was no methane or less than 5 ppbv of methane at the landing site.^{[67][68][69][70][71]} On 2013, NASA scientists again reported no detection of methane beyond a baseline.^{[72][73][74]} But on 2014, NASA reported that the Curiosity rover detected a tenfold increase ('spike') in methane in the atmosphere around it in late 2013 and early 2014. Four measurements taken over two months in this period averaged 7.2 ppbv, implying that Mars is episodically producing or releasing methane from an unknown source.^[75] Before and after that, readings averaged around one-tenth that level.^{[8][9][75]}



Methane measurements in the atmosphere of Mars by the *Curiosity* rover.

The Indian Mars Orbiter Mission, which entered orbit around Mars on 24 September 2014, is equipped with a Fabry–Pérot interferometer to measure atmospheric methane at a level of several ppb, which is systematically gathering data as of September 2015.^{[76][77]} The ExoMars Trace Gas Orbiter, which entered orbit on 19 October 2016, will further study the methane, as well as its decomposition products such as formaldehyde and methanol.^{[61][78][79][80]}

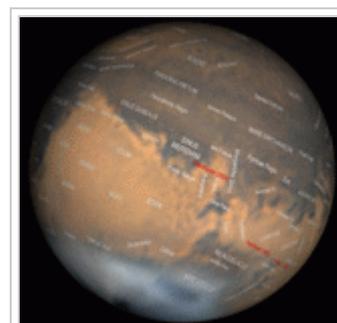
Sulfur dioxide

Sulfur dioxide in the atmosphere is thought to be a tracer of current volcanic activity. It has become especially interesting due to the long-standing controversy of methane

on Mars. If methane on Mars were being produced by volcanoes (as it is in part on Earth) we would expect to find sulfur dioxide in large quantities. Several teams have searched for sulfur dioxide on Mars using the NASA Infrared Telescope Facility. No sulfur dioxide was detected in these studies, but they were able to place stringent upper limits on the atmospheric concentration of 0.2 ppb.^{[62][63]} In March 2013, a team led by scientists at NASA Goddard Space Flight Center reported a detection of SO₂ in Rocknest soil samples analyzed by the *Curiosity* rover.^[81]

Ozone

As reported by the European Space Agency (ESA) on September 29, 2013, a new comparison of spacecraft data with computer models explains how global atmospheric circulation creates a layer of ozone (O₃) above Mars's southern pole in winter. Ozone was most likely difficult to detect on Mars because its concentration is typically 300 times lower than on Earth, although it varies greatly with location and time. The SPICAM—an UV/IR spectrometer—on board Mars Express has shown the presence of two distinct ozone layers at low-to-mid latitudes. These comprise a persistent, near-surface layer below an altitude of 30 km, a separate layer that is only present in northern spring and summer with an altitude varying from 30 to 60 km, and another separate layer that exists 40–60 km above the southern pole in winter, with no counterpart above the Mars's north pole. This third ozone layer shows an abrupt decrease in elevation between 75 and 50 degrees south. SPICAM detected a gradual increase in ozone concentration at 50 km until midwinter, after which it slowly decreased to very low concentrations, with no layer detectable above 35 km. The reporting scientists think that the observed polar ozone layers are the result of the same atmospheric circulation pattern that creates a distinct oxygen emission identified in the polar night and also present in Earth's atmosphere. This circulation takes the form of a huge Hadley cell in which warmer air rises and travels toward the south pole before cooling and sinking at higher latitudes. Mars is on a quite elliptical orbit and has a large axial tilt, which causes extreme seasonal variations in temperature amongst the northern and southern hemispheres. Mars's temperature difference greatly influences the amount of water vapor in the atmosphere, because warmer air can contain more moisture. This, in turn, affects the production of ozone-destroying hydrogen radicals.



Rotation of Mars near opposition. Ecliptic south is up.

Oxygen

In early 2016, Stratospheric Observatory for Infrared Astronomy (SOFIA) detected atomic oxygen in the atmosphere of Mars.^[82] This was the first time in forty years it was detected, the last time being the Viking and Mariner missions in the 1970s.^[82]

Potential for use by humans

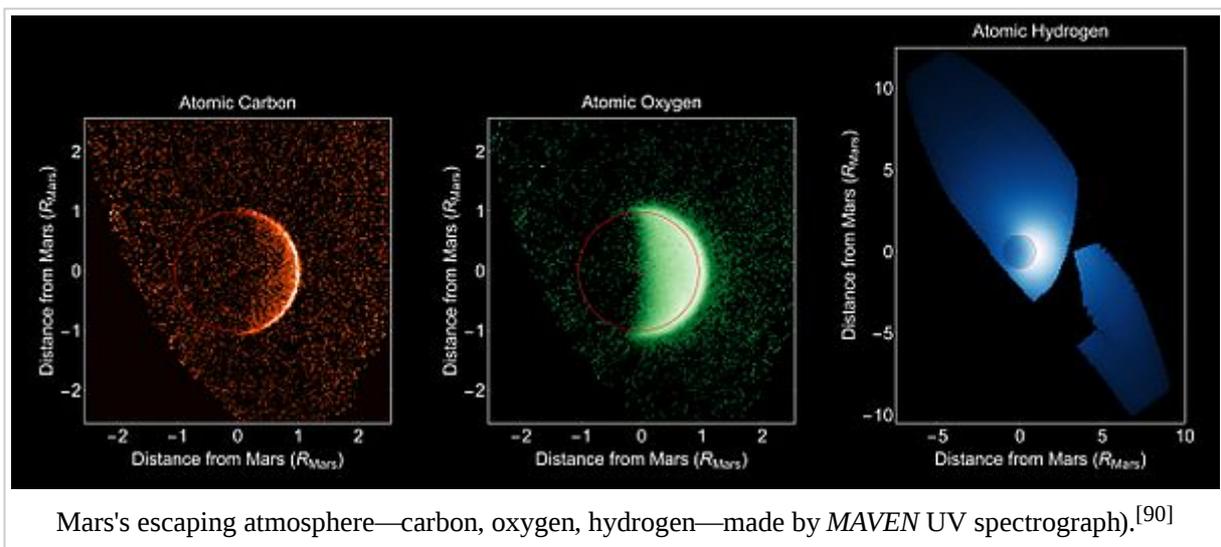
The atmosphere of Mars is a resource of known composition available at any landing site on Mars. It has been proposed that human exploration of Mars could use carbon dioxide (CO_2) from the Martian atmosphere to make rocket fuel for the return mission. Mission studies that propose using the atmosphere in this way include the Mars Direct proposal of Robert Zubrin and the NASA Design reference mission study. Two major chemical pathways for use of the carbon dioxide are the Sabatier reaction, converting atmospheric carbon dioxide along with additional hydrogen (H_2), to produce methane (CH_4) and oxygen (O_2), and electrolysis, using a zirconia solid oxide electrolyte to split the carbon dioxide into oxygen (O_2) and carbon monoxide (CO).

History

Mars's atmosphere is thought to have changed over the course of the planet's lifetime, with evidence suggesting the possibility that Mars had large oceans a few billion years ago.^[83] As stated in the Mars ocean hypothesis, atmospheric pressure on the present-day Martian surface only exceeds that of the triple point of water (6.11 hectopascals (0.0886 psi)) in the lowest elevations; at higher elevations water can exist only in solid or vapor form. Annual mean temperatures at the surface are currently $< 210 \text{ K}$ ($-63 \text{ }^\circ\text{C}$; $-82 \text{ }^\circ\text{F}$), significantly lower than that needed to sustain liquid water. However, early in its history Mars may have had conditions more conducive to retaining liquid water at the surface. In 2013, scientists published that Mars once had "oxygen-rich" atmosphere billions of years ago.^{[84][85]}

Possible causes for the depletion of a previously thicker Martian atmosphere include:

- Gradual erosion of the atmosphere by solar wind. On 5 November 2015, NASA announced that data from MAVEN shows that the erosion of Mars' atmosphere increases significantly during solar storms. This shift took place between about 4.2 to 3.7 billion years ago, as the shielding effect of the global magnetic field was lost when the planet's internal dynamo cooled.^{[13][14][86][87]}
- Catastrophic collision by a body large enough to blow away a significant percentage of the atmosphere;^[88]
- Mars' low gravity allowing the atmosphere to "blow off" into space by Jeans escape.^[89]



Images



Mars's thin atmosphere, visible on the horizon.



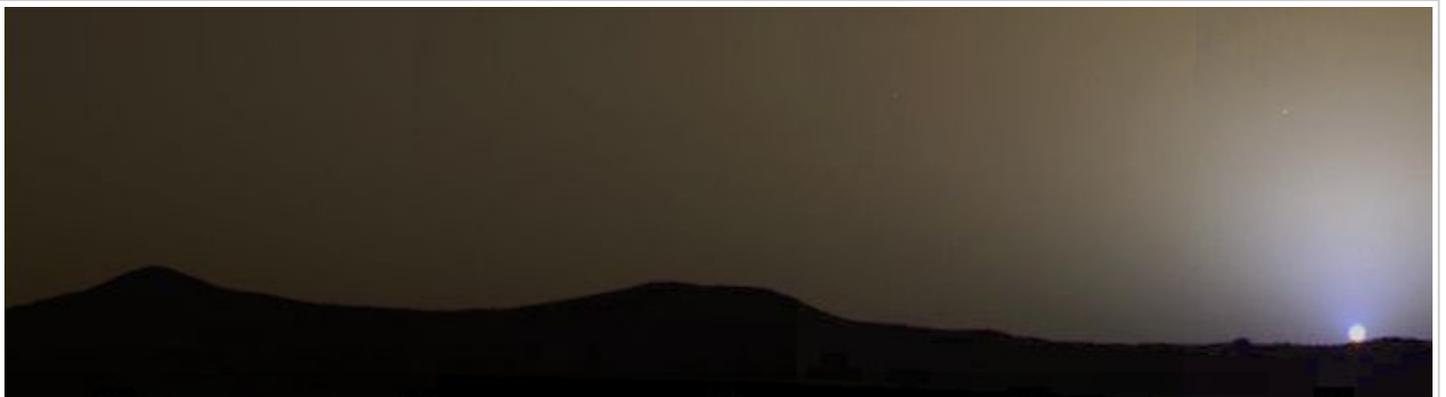
Mars Pathfinder – Martian sky with water ice clouds.



A storm front moves in

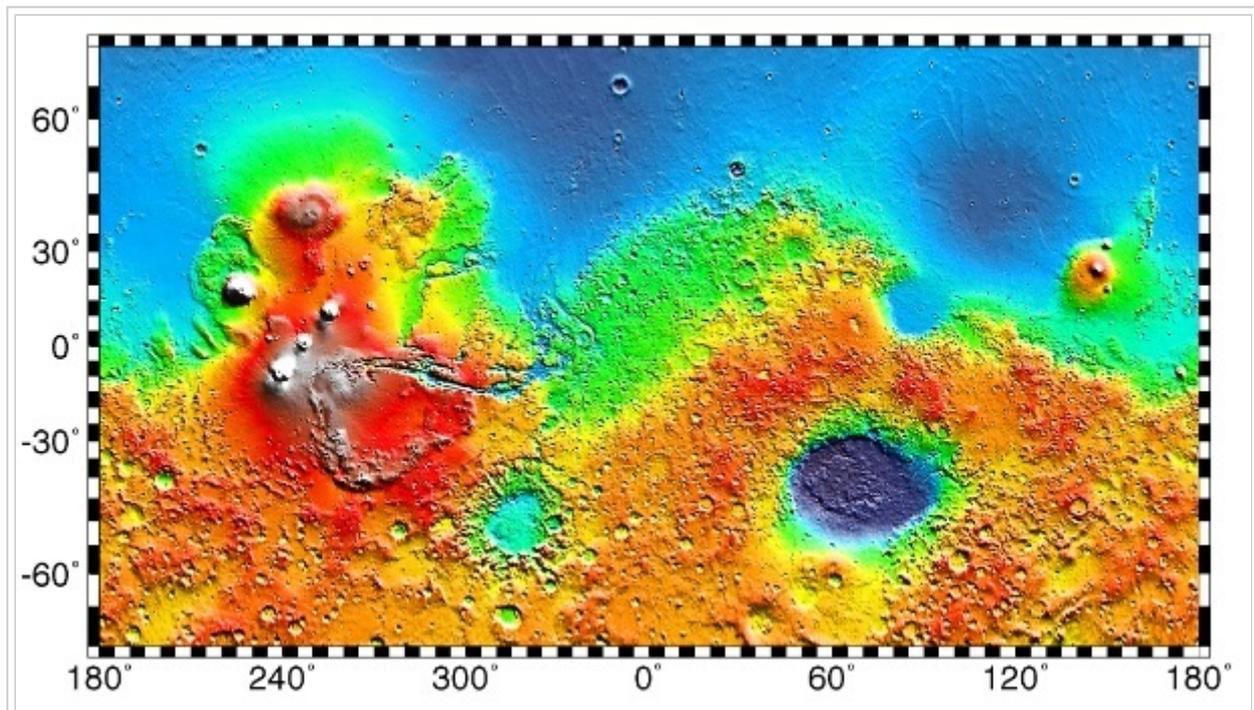


Martian sunset by Spirit rover at Gusev crater (May, 2005).



Martian sunset by Pathfinder at Ares Vallis (July, 1997).

Interactive Mars map



Interactive imagemap of the global topography of Mars. Hover your mouse to see the names of over 25 prominent geographic features, and click to link to them. Coloring of the base map indicates relative elevations, based on data from the Mars Orbiter Laser Altimeter on NASA's Mars Global Surveyor. Reds and pinks are higher elevation (+3 km to +8 km); yellow is 0 km; greens and blues are lower elevation (down to -8 km). Whites (>+12 km) and browns (>+8 km) are the highest elevations. Axes are latitude and longitude; Poles are not shown.

(See also: *Mars Rovers map*) (view • discuss)

See also

- Climate of Mars
- In situ resource utilization
- Life on Mars
- Mars MetNet – proposed surface observation network
- Mars regional atmospheric modeling system
- MAVEN orbiter
- Seasonal flows on warm Martian slopes
- Terraforming of Mars

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Atmosphere of Venus

From Wikipedia, the free encyclopedia

The **atmosphere of Venus** is the layer of gases surrounding Venus. It is composed primarily of carbon dioxide and is much denser and hotter than that of Earth. The temperature at the surface is 740 K (467 °C, 872 °F), and the pressure is 93 bar (9.3 MPa), roughly the pressure found 900 m (3,000 ft) underwater on Earth.^[1] The Venusian atmosphere supports opaque clouds made of sulfuric acid, making optical Earth-based and orbital observation of the surface impossible. Information about the topography has been obtained exclusively by radar imaging.^[1] Aside from carbon dioxide, the other main component is nitrogen. Other chemical compounds are present only in trace amounts.^[1]

Mikhail Lomonosov was the first person to hypothesize the existence of an atmosphere on Venus based on his observation of the transit of Venus of 1761 in a small observatory near his house in Saint Petersburg, Russia.^[3]

Aside from the very surface layers, the atmosphere is in a state of vigorous circulation.^[4] The upper layer of troposphere exhibits a phenomenon of super-rotation, in which the atmosphere circles the planet in just four Earth days, much faster than the planet's sidereal day of 243 days. The winds supporting super-rotation blow at a speed of 100 m/s (~360 km/h or 220 mph)^[4] or more. Winds move at up to 60 times the speed of the planet's rotation, while Earth's fastest winds are only 10% to 20% rotation speed.^[5] On the other hand, the wind speed becomes increasingly slower as the elevation from the surface decreases, with the breeze barely reaching the speed of 10 km/h (2.8 m/s) on the surface.^[6] Near the poles are anticyclonic structures called polar vortices. Each vortex is double-eyed and shows a characteristic S-shaped pattern of clouds.^[7] Above there is an intermediate layer of mesosphere which separates the troposphere from the thermosphere.^{[4][2]} The thermosphere is also characterized by strong circulation, but very different in its nature - the gases heated and partially ionized by sunlight in the sunlit hemisphere migrate to the dark hemisphere where they recombine and downwell.^[2]

Unlike Earth, Venus lacks a magnetic field. Its ionosphere separates the atmosphere from outer space and the solar wind. This ionised layer excludes the solar magnetic field, giving Venus a distinct magnetic environment. This is considered Venus's *induced magnetosphere*. Lighter gases, including water vapour, are continuously blown away by the solar wind through the induced magnetotail.^[4] It is speculated that the atmosphere of Venus up to around 4 billion years ago was more like that of the Earth with liquid water on the surface. A runaway greenhouse effect may have been caused by the evaporation of the surface water and subsequent rise of the levels of other greenhouse gases.^{[8][9]}

Atmosphere of Venus



Cloud structure in Venus's atmosphere in 1979, revealed by ultraviolet observations from Pioneer Venus Orbiter

General information^[1]

Height	250 km
Average surface pressure	93 bar or 9.3 MPa
Mass	4.8×10^{20} kg

Composition^{[1][2]}

Carbon dioxide	96.5 %
Nitrogen	3.5 %
Sulfur dioxide	150 ppm
Argon	70 ppm
Water vapour	20 ppm
Carbon monoxide	17 ppm
Helium	12 ppm
Neon	7 ppm
Hydrogen chloride	0.1–0.6 ppm
Hydrogen fluoride	0.001–0.005 ppm

Despite the harsh conditions on the surface, the atmospheric pressure and temperature at about 50 km to 65 km above the surface of the planet is nearly the same as that of the Earth, making its upper atmosphere the most Earth-like area in the Solar System, even more so than the surface of Mars. Due to the similarity in pressure and temperature and the fact that breathable air (21% oxygen, 78% nitrogen) is a lifting gas on Venus in the same way that helium is a lifting gas on Earth, the upper atmosphere has been proposed as a location for both exploration and colonization.^[10]

On January 29, 2013, ESA scientists reported that the ionosphere of the planet Venus streams outwards in a manner similar to "the ion tail seen streaming from a comet under similar conditions."^{[11][12]}

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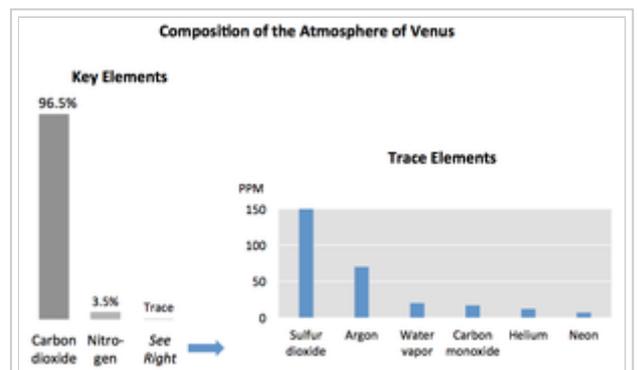
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Structure and composition

Composition

The atmosphere of Venus is composed of 96.5% carbon dioxide, 3.5% nitrogen, and traces of other gases, most notably sulfur dioxide.^[13] The amount of nitrogen in the atmosphere is relatively small compared to the amount of carbon dioxide, but because the atmosphere is so much thicker than that on Earth, its total nitrogen content is roughly four times higher than Earth's, even though on Earth nitrogen makes up about 78% of the atmosphere.^{[1][14]}

The atmosphere contains a range of interesting compounds in small quantities, including some based on hydrogen, such as hydrogen chloride (HCl) and hydrogen fluoride (HF). There is carbon monoxide, water vapour and atomic oxygen as well.^{[2][4]} Hydrogen is in relatively short supply in the Venusian atmosphere. A large amount of the planet's hydrogen is theorised to have been lost to space,^[15]

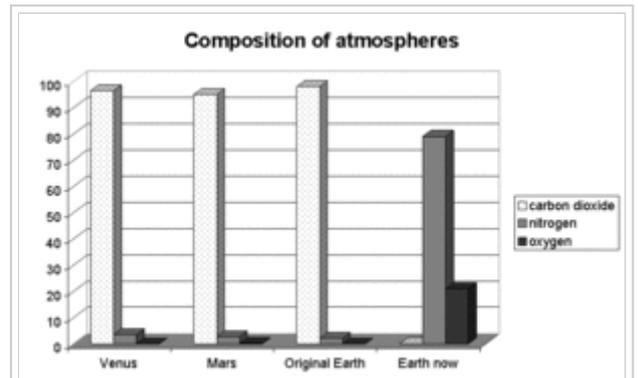


Composition of the atmosphere of Venus. The chart on the right is an expanded view of the trace elements that all together do not even make up a tenth of a percent.

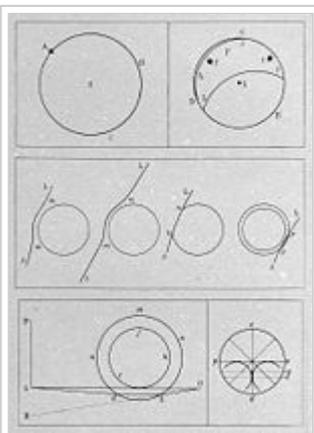
with the remainder being mostly bound up in sulfuric acid (H₂SO₄) and hydrogen sulfide (H₂S). The loss of significant amounts of hydrogen is proven by a very high D–H ratio measured in the Venusian atmosphere.^[4] The ratio is about 0.015–0.025, which is 100–150 times higher than the terrestrial value of 1.6×10^{-4} .^{[2][16]} According to some measurements, in the upper atmosphere of Venus D/H ratio is 1.5 higher than in the bulk atmosphere.^[2]

Troposphere

The atmosphere is divided into a number of sections depending on altitude. The densest part of the atmosphere, the troposphere, begins at the surface and extends upwards to 65 km. At the furnace-like surface the winds are slow,^[1] but at the top of the troposphere the temperature and pressure reaches Earth-like levels and clouds pick up speed to 100 m/s.^{[4][17]}



Comparison of Atmosphere Compositions – Venus, Mars, Earth (past and present).



1761 drawing by Mikhail Lomonosov in his work on discovery of atmosphere of Venus

The atmospheric pressure at the surface of Venus is about 92 times that of the Earth, similar to the pressure found 900 m (3,000 ft) below the surface of the ocean. The

atmosphere has a mass of 4.8×10^{20} kg, about 93 times the mass of the Earth's total atmosphere. The density of the air at the surface is 67 kg/m^3 , which is 6.5% that of liquid water on Earth.^[1] The pressure found on Venus's surface is high enough that the carbon dioxide is technically no longer a gas, but a supercritical fluid. This supercritical carbon dioxide forms a kind of sea that covers the entire surface of Venus. This sea of supercritical carbon dioxide transfers heat very efficiently, buffering the temperature changes between night and day (which last 56 terrestrial days).^[18]

The large amount of CO₂ in the atmosphere together with water vapour and sulfur dioxide create a strong greenhouse effect, trapping solar energy and raising the surface temperature to around 740 K (467 °C),^[14] hotter than any other planet in the Solar System, even that of Mercury despite being located farther out from the Sun and receiving only 25% of the solar energy (per unit area) Mercury does. The average temperature on the surface is above the melting points of lead (600 K, 327 °C), tin (505 K, 232 °C), and zinc (693 K, 420 °C). The thick troposphere also makes the difference in temperature between the day and night side small, even though the slow retrograde rotation of the planet causes a single solar day to last 116.5 Earth days. The surface of Venus spends 58.3 days of darkness before the sun rises again behind the clouds.^[1]

The troposphere on Venus contains 99% of the atmosphere by mass. Ninety percent of the atmosphere of Venus is within 28 km of the surface; by comparison, 90% of the atmosphere of Earth is within 10 km of the surface. At a height of 50 km the atmospheric pressure is approximately equal to that at the surface of Earth.^[20] On the night side of Venus clouds can still be found at 80 km above the surface.^[21]

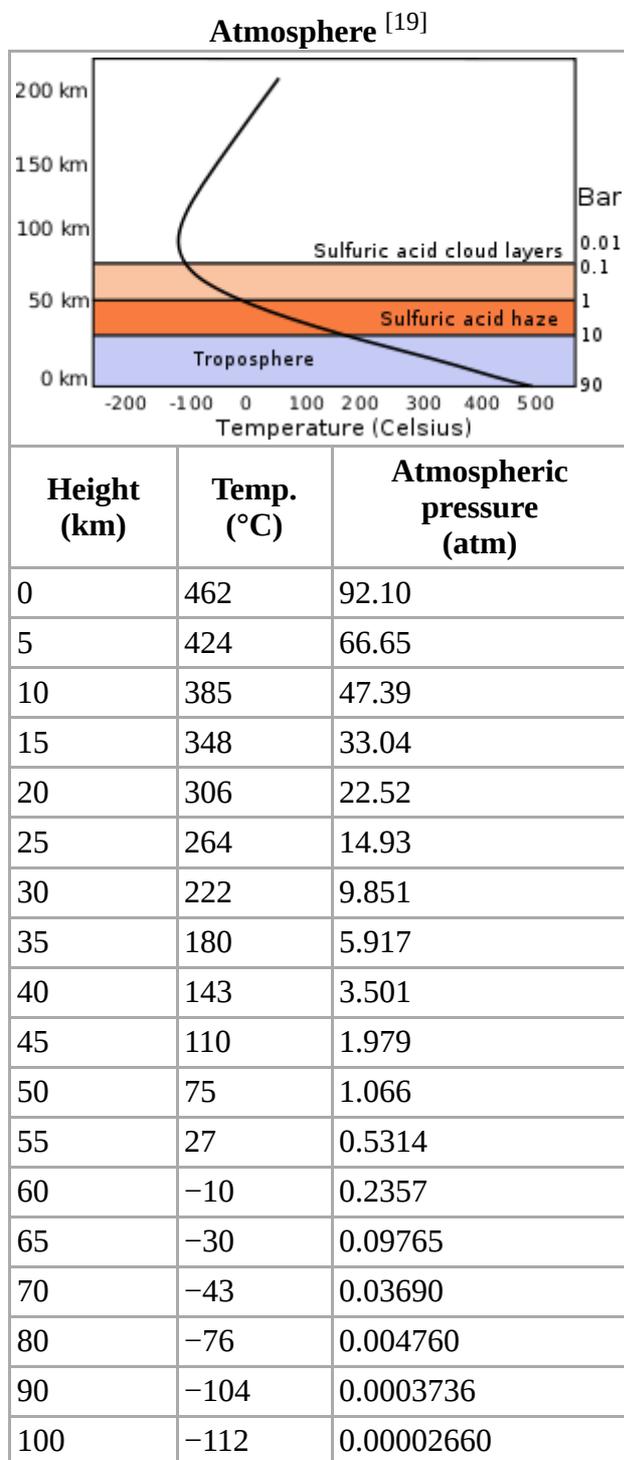
The altitude of the troposphere most similar to Earth is near the tropopause—the boundary between troposphere and mesosphere. It is located slightly above 50 km.^[17] According to measurements by the *Magellan* and *Venus Express* probes, the altitude from 52.5 to 54 km has a temperature between 293 K (20 °C) and 310 K (37 °C), and the altitude at 49.5 km above the surface is where the pressure becomes the same as Earth at sea level.^{[17][22]} As manned ships sent to Venus would be able to compensate for differences in temperature to a certain extent, anywhere from about 50 to 54 km or so above the surface would be the easiest altitude in which

to base an exploration or colony, where the temperature would be in the crucial "liquid water" range of 273 K (0 °C) to 323 K (50 °C) and the air pressure the same as habitable regions of Earth.^{[10][23]} As CO₂ is heavier than air, the colony's air (nitrogen and oxygen) could keep the structure floating at that altitude like a dirigible.

Circulation

The circulation in Venus's troposphere follows the so-called cyclostrophic approximation.^[4] Its windspeeds are roughly determined by the balance of the pressure gradient and centrifugal forces in almost purely zonal flow. In contrast, the circulation in the Earth's atmosphere is governed by the geostrophic balance.^[4] Venus's windspeeds can be directly measured only in the upper troposphere (tropopause), between 60–70 km, altitude, which corresponds to the upper cloud deck.^[24] The cloud motion is usually observed in the ultraviolet part of the spectrum, where the contrast between clouds is the highest.^[24] The linear wind speeds at this level are about 100 ± 10 m/s at lower than 50° latitude. They are retrograde in the sense that they blow in the direction of the retrograde rotation of the planet.^[24] The winds quickly decrease towards the higher latitudes, eventually reaching zero at the poles. Such strong cloud-top winds cause a phenomenon known as the super-rotation of the atmosphere.^[4] In other words, these high-speed winds circle the whole planet faster than the planet itself rotates.^[23] The super-rotation on Venus is differential, which means that the equatorial troposphere super-rotates more slowly than the troposphere at the midlatitudes.^[24] The winds also have a strong vertical gradient. They decline deep in the troposphere with the rate of 3 m/s per km.^[4] The winds near the surface of Venus are much slower than that on Earth. They actually move at only a few kilometres per hour (generally less than 2 m/s and with an average of 0.3 to 1.0 m/s), but due to the high density of the atmosphere at the surface, this is still enough to transport dust and small stones across the surface, much like a slow-moving current of water.^{[1][25]}

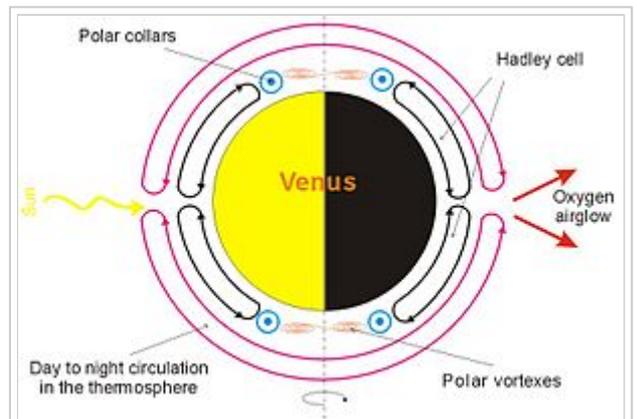
All winds on Venus are ultimately driven by convection.^[4] Hot air rises in the equatorial zone, where solar heating is concentrated, and flows to the poles. Such an almost-planetwide overturning of the troposphere is called Hadley circulation.^[4] However, the meridional air motions are much slower than zonal winds. The poleward limit of the planet wide Hadley cell on Venus is near $\pm 60^\circ$ latitudes.^[4] Here air starts to descend and returns to the equator below the clouds. This interpretation is supported by the distribution of the carbon monoxide, which is also concentrated in the vicinity of $\pm 60^\circ$ latitudes.^[4] Poleward of the Hadley cell a different pattern of circulation is observed. In the latitude range 60° – 70° cold polar collars exist.^{[4][7]} They are characterised by temperatures about 30–40 K lower than in the upper troposphere at nearby latitudes.^[7] The lower temperature is probably caused by the upwelling of the air in them and by the resulting adiabatic cooling.^[7] Such an interpretation is supported by the denser and higher clouds in the collars. The clouds lie at



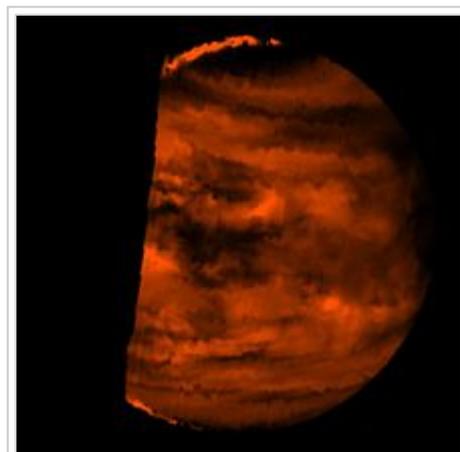
70–72 km altitude in the collars—about 5 km higher than at the poles and low latitudes.^[4] A connection may exist between the cold collars and high speed midlatitude jets in which winds blow as fast as 140 m/s. Such jets are a natural consequence of the Hadley-type circulation and should exist on Venus between 55–60° latitude.^[24]

Odd structures known as polar vortices lie within the cold polar collars.^[4] They are giant hurricane-like storms four times larger than their terrestrial analogs. Each vortex has two "eyes"—the centres of rotation, which are connected by distinct S-shaped cloud structures. Such double eyed structures are also called polar dipoles.^[7] Vortexes rotate with the period of about 3 days in the direction of general super-rotation of the atmosphere.^[7] The linear wind speeds are 35–50 m/s near their outer edges and zero at the poles.^[7] The temperature at the cloud-tops in the polar

vortexes are much higher than in the nearby polar collars reaching 250 K (–23 °C).^[7] The conventional interpretation of the polar vortexes is that they are anticyclones with downwelling in the centre and upwelling in the cold polar collars.^[7] This type of circulation resembles the winter polar anticyclonic vortexes on Earth, especially the one found over Antarctica. The observations in the various infrared atmospheric windows indicate that the anticyclonic circulation observed near the poles may penetrate as deep as to 50 km altitude, i.e. to the base of the clouds.^[7] The polar upper troposphere and mesosphere are extremely dynamic; large bright clouds may appear and disappear over the space of a few hours. One such event was observed by Venus Express between 9 and 13 January 2007, when the south polar region became brighter by 30%.^[24] This event was probably caused by an injection of sulfur dioxide into the mesosphere, which then condensed forming a bright haze.^[24] The two eyes in the vortexes have yet to be explained.^[26]



Meridional (north–south) component of the atmospheric circulation in the atmosphere of Venus. Note that the meridional circulation is much lower than the zonal circulation, which transports heat between the day and night sides of the planet



False colour near infrared (2.3 μm) image of the deep atmosphere of Venus obtained by *Galileo*. The dark spots are clouds silhouetted against the very hot lower atmosphere emitting thermal infrared radiation.

The first vortex on Venus was discovered at the north pole by the Pioneer Venus mission in 1978.^[27] A discovery of the second large 'double-eyed' vortex at the south pole of Venus was made in the summer of 2006 by Venus Express, which came with no surprise.^[26]

Upper atmosphere and ionosphere

The mesosphere of Venus extends from 65 km to 120 km in height, and the thermosphere begins at approximately 120 km, eventually reaching the upper limit of the atmosphere (exosphere) at about 220 to 350 km.^[17] The exosphere is the altitude at which the atmosphere becomes collisionless.

The mesosphere of Venus can be divided into two layers: the lower one between 62–73 km^[28] and the upper one between 73–95 km.^[17] In the first layer the temperature is nearly constant at 230 K (–43 °C). This layer coincides with the upper cloud deck. In the second layer temperature starts to decrease again reaching about 165 K (–108 °C) at the altitude of 95 km, where mesopause begins.^[17] It is the coldest part

of the Venusian dayside atmosphere.^[2] In the dayside mesopause, which serves as a boundary between the mesosphere and thermosphere and is located between 95–120 km, temperature increases to a constant—about

300–400 K (27–127 °C)—value prevalent in the thermosphere.^[2] In contrast, the nightside Venusian thermosphere is the coldest place on Venus with temperature as low as 100 K (–173 °C). It is even called a cryosphere.^[2]

The circulation patterns in the upper mesosphere and thermosphere of Venus are completely different from those in the lower atmosphere.^[2] At altitudes 90–150 km the Venusian air moves from the dayside to nightside of the planet, with upwelling over sunlit hemisphere and downwelling over dark hemisphere. The downwelling over the nightside causes adiabatic heating of the air, which forms a warm layer in the nightside mesosphere at the altitudes 90–120 km.^{[4][2]} The temperature of this layer—230 K (–43 °C) is far higher than the typical temperature found in the nightside thermosphere—100 K (–173 °C).^[2] The air circulated from the dayside also carries oxygen atoms, which after recombination form excited molecules of oxygen in the long-lived singlet state ($^1\Delta_g$), which then relax and emit infrared radiation at the wavelength 1.27 μm . This radiation from the altitude range 90–100 km is often observed from the ground and spacecraft.^[29] The nightside upper mesosphere and thermosphere of Venus is also the source of non-local thermodynamic equilibrium emissions of CO_2 and nitric oxide molecules, which are responsible for the low temperature of the nightside thermosphere.^[29]

The Venus Express probe has shown through stellar occultation that the atmospheric haze extends much further up on the night side than the day side. On the day side the cloud deck has a thickness of 20 km and extends up to about 65 km, whereas on the night side the cloud deck in the form of a thick haze reaches up to 90 km in altitude—well into mesosphere, continuing even further to 105 km as a more transparent haze.^[21] In 2011, the spacecraft discovered that Venus has a thin ozone layer at an altitude of 100 km.^[30]

Venus has an extended ionosphere located at altitudes 120–300 km.^[17] The ionosphere almost coincides with the thermosphere. The high levels of the ionization are maintained only over the dayside of the planet. Over the nightside the concentration of the electrons is almost zero.^[17] The ionosphere of Venus consists of three layers: v1 between 120 and 130 km, v2 between 140 and 160 km and v3 between 200 and 250 km.^[17] There may be an additional layer near 180 km. The maximum electron volume density (number of electrons in a unit of volume) $3 \times 10^{11} \text{ m}^{-3}$ is reached in the v2 layer near the subsolar point.^[17] The upper boundary of the ionosphere—ionopause is located at altitudes 220–375 km and separates the plasma of the planetary origin from that of the induced magnetosphere.^{[31][32]} The main ionic species in the v1 and v2 layers is O_2^+ ion, whereas the v3 layer consists of O^+ ions.^[17] The ionospheric plasma is observed to be in motion; solar photoionization on the dayside, and ion recombination on the nightside, are the processes mainly responsible for accelerating the plasma to the observed velocities. The plasma flow appears to be sufficient to maintain the nightside ionosphere at or near the observed median level of ion densities.^[33]

Induced magnetosphere

Venus is known not to have a magnetic field.^{[31][32]} The reason for its absence is not clear, but is probably related to the planet's slow rotation or the lack of convection in the mantle. Venus only has an induced magnetosphere formed by the Sun's magnetic field carried by the solar wind.^[31] This process can be understood as the field lines wrapping around an obstacle—Venus in this case. The induced magnetosphere of Venus has a bow shock, magnetosheath, magnetopause and magnetotail with the current sheet.^{[31][32]}

At the subsolar point the bow shock stands 1900 km ($0.3 R_v$, where R_v is the radius of Venus) above the surface of Venus. This distance was measured in 2007 near the solar activity minimum.^[32] Near the solar activity maximum it can be several times further from the planet.^[31] The magnetopause is located at the altitude of 300 km.^[32] The upper boundary of the ionosphere (ionopause) is near 250 km. Between the magnetopause and ionopause there exists a magnetic barrier—a local enhancement of the magnetic field, which prevents solar

plasma from penetrating deeper into the Venusian atmosphere, at least near solar activity minimum. The magnetic field in the barrier reaches up to 40 nT.^[32] The magnetotail continues up to ten radiuses from the planet. It is the most active part of the Venusian magnetosphere. There are reconnection events and particle acceleration in the tail. The energies of electrons and ions in the magnetotail are around 100 eV and 1000 eV respectively.^[34]

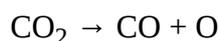
Due to the lack of the intrinsic magnetic field on Venus, the solar wind penetrates relatively deep into the planetary exosphere and causes substantial atmosphere loss.^[35] The loss happens mainly via the magnetotail. Currently the main ion types being lost are O^+ , H^+ and He^+ . The ratio of hydrogen to oxygen losses is around 2 (i.e. almost stoichiometric) indicating the ongoing loss of water.^[34]

Clouds

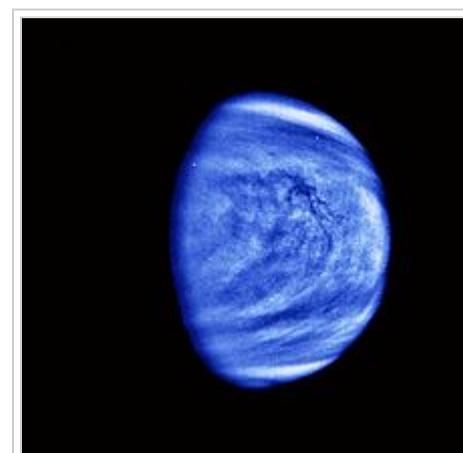
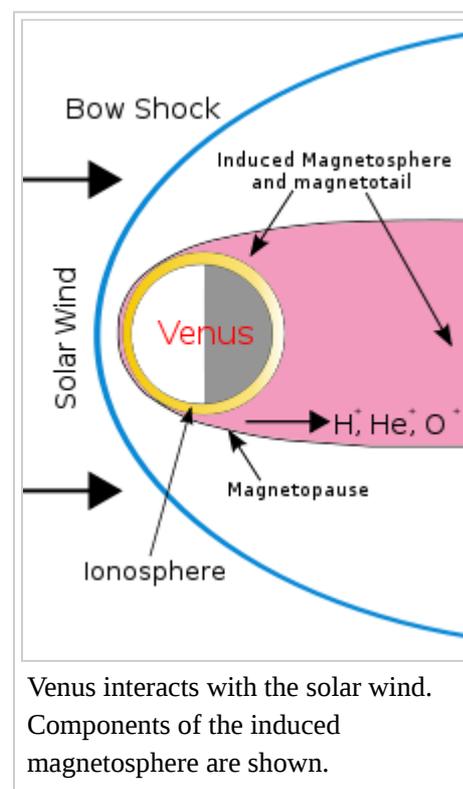
Venusian clouds are thick and are composed of sulfur dioxide and droplets of sulfuric acid.^[36] These clouds reflect about 75%^[37] of the sunlight that falls on them, which is what obscures the surface of Venus from optical imaging.^[1] The geometric albedo, a common measure of reflectivity, is the highest of any planet in the Solar System. This high reflectivity potentially enables any probe exploring the cloud tops sufficient solar energy such that solar cells can be fitted anywhere on the craft.^[38] The density of the clouds is highly variable with the densest layer at about 48.5 km, reaching 0.1 g/m^3 similar to the lower range of cumulonimbus storm clouds on Earth.^[39]

The cloud cover is such that typical surface light levels are similar to a partly cloudy day on Earth, around 5000–10000 lux. The equivalent visibility is about three kilometers, but this will likely vary with the wind conditions. Little to no solar energy could conceivably be collected by solar panels on a surface probe. In fact, due to the thick, highly reflective cloud cover, the total solar energy received by the surface of the planet is less than that of the Earth.

Sulfuric acid is produced in the upper atmosphere by the Sun's photochemical action on carbon dioxide, sulfur dioxide, and water vapour.^[40] Ultraviolet photons of wavelengths less than 169 nm can photodissociate carbon dioxide into carbon monoxide and atomic oxygen. Atomic oxygen is highly reactive; when it reacts with sulfur dioxide, a trace component of the Venusian atmosphere, the result is sulfur trioxide, which can combine with water vapour, another trace component of Venus's atmosphere, to yield sulfuric acid.



Surface level humidity is less than 0.1%.^[41] Venus's sulfuric acid rain never reaches the ground, but is evaporated by the heat before reaching the surface in a phenomenon known as virga.^[42] It is theorized that early volcanic activity released sulfur into the atmosphere and the high temperatures prevented it from being trapped into solid compounds on the surface as it was on the Earth.^[43]



Photograph taken by the unmanned *Galileo* spaceprobe en route to Jupiter in 1990 during a Venus flyby. Smaller-scale cloud features have been emphasized and a bluish hue has been applied to show that it was taken through a violet filter.

In 2009 a prominent bright spot in the atmosphere was noted by an amateur astronomer and photographed by *Venus Express*. Its cause is currently unknown, with surface volcanism advanced as a possible explanation.^[44]

Lightning

The clouds of Venus are capable of producing lightning much like the clouds on Earth.^[45] The existence of lightning had been controversial outside of the former Soviet Union since the first suspected bursts were detected by the Soviet *Venera* probes. However, in 2006–2007 *Venus Express* was reported to detect whistler mode waves, which were attributed to lightning. Their intermittent appearance indicates a pattern associated with weather activity. The lightning rate is at least half of that on Earth.^[45]

The sulfuric acid droplets can be highly electrically charged, and so they offer the potential for lightning.^[40] Venus' small induced magnetosphere provides negligible protection to the atmosphere against cosmic radiation. This radiation may result in cloud-to-cloud lightning discharges.^[46]

Throughout the 1980s, it was thought that the cause of the glow was lightning on Venus.^[47] The Soviet *Venera* 9 and 10 orbiters obtained optical and electromagnetic evidence of lightning on Venus.^[48]

^[49] Also, the *Pioneer Venus Orbiter* recorded visible airglow at Venus in 1978 strong enough to saturate its star sensor.^[48] The European Space Agency's *Venus Express* in 2007 detected whistler waves further confirming the occurrence of lightning on Venus.^{[50][51]}

Possibility of life

Due to the harsh conditions on the surface, little of the planet has been explored; in addition to the fact that life as currently understood may not necessarily be the same in other parts of the universe, the extent of the tenacity of life on Earth itself has not yet been shown. Creatures known as extremophiles exist on Earth, preferring extreme habitats. Thermophiles and hyperthermophiles thrive at temperatures reaching above the boiling point of water, acidophiles thrive at a pH level of 3 or below, polyextremophiles can survive a varied number of extreme conditions, and many other types of extremophiles exist on Earth.^[52]

The surface temperature of Venus (over 450 °C) is far beyond the extremophile range, which extends only tens of degrees beyond 100 °C. However, the lower temperature of the cloud tops means that life could plausibly exist. It has been proposed that life on Venus could exist there, the same way that bacteria have been found living and reproducing in clouds on Earth.^[53] Any such bacteria living in the cloud tops, however, would have to be hyper-acidiphilic, due to the concentrated sulfuric acid environment. Microbes in the thick, cloudy atmosphere could be protected from solar radiation by the sulfur compounds in the air.^[52] The solar wind may provide a mechanism for the transfer of such microbiota from Venus to Earth.^[54]

The Venusian atmosphere has been found to be sufficiently out of equilibrium as to require further investigation.^[52] Analysis of data from the *Venera*, *Pioneer*, and *Magellan* missions has found the chemicals hydrogen sulfide (H₂S) and sulfur dioxide (SO₂) together in the upper atmosphere, as well as carbonyl sulfide (OCS). The first two gases react with each other, implying that something must produce them. Carbonyl sulfide is difficult to produce inorganically, but it is present in the Venusian atmosphere.^[53] However, the planet's volcanism could explain the presence of carbonyl sulfide.^[53] In addition, one of the early *Venera* probes detected large amounts of toxic chlorine just below the Venusian cloud deck.^[55]

It has been proposed that microbes at this level could be soaking up ultraviolet light from the Sun as a source of energy, which could be a possible explanation for dark patches seen on UV images of the planet.^{[56][57]} Large, non-spherical cloud particles have also been detected in the cloud decks. In 2012, the abundance and vertical distribution of these unknown ultraviolet absorbers in the Venusian atmosphere have been investigated from

analysis of Venus Monitoring Camera images,^[58] but their composition is still unknown.^[52] In 2016, disulfur dioxide has been identified as a possible candidate for causing the so far unknown UV absorption of the Venusian atmosphere.^[59]

Evolution

Through studies of the present cloud structure and geology of the surface, combined with the fact that the luminosity of the Sun has increased by 25% since around 3.8 billion years ago,^[60] it is thought that the atmosphere of Venus up to around 4 billion years ago was more like that of Earth with liquid water on the surface. The runaway greenhouse effect may have been caused by the evaporation of the surface water and the rise of the levels of greenhouse gases that followed. Venus's atmosphere has therefore received a great deal of attention from those studying climate change on Earth.^[8]

There are no geologic forms on the planet to suggest the presence of water over the past billion years. However, there is no reason to suppose that Venus was an exception to the processes that formed Earth and gave it its water during its early history, possibly from the original rocks that formed the planet or later on from comets. The common view among research scientists is that water would have existed for about 600 million years on the surface before evaporating, though some such as David Grinspoon believe that up to 2 billion years could also be plausible.^[61]

The early Earth during the Hadean eon is believed by most scientists to have had a Venus-like atmosphere, with roughly 100 bar of CO₂ and a surface temperature of 230 °C, and possibly even sulfuric acid clouds, until about 4.0 billion years ago, by which time plate tectonics were in full force and together with the early water oceans, removed the CO₂ and sulfur from the atmosphere.^[62] Early Venus would thus most likely have had water oceans like the Earth, but any plate tectonics would have ended when Venus lost its oceans. Its surface is estimated to be about 500 million years old, so it would not be expected to show evidence of plate tectonics.^[63]

Observations and measurement from Earth



Venus transits the face of the Sun on June 8, 2004, providing valuable information on the upper atmosphere through spectroscopic measurements from Earth

In 1761, Russian polymath Mikhail Lomonosov observed an arc of light surrounding the part of Venus off the Sun's disc at the beginning of the egress phase of the transit and concluded that Venus has an atmosphere.^{[64][65]} In 1940, Rupert Wildt calculated that the amount of CO₂ in the Venusian atmosphere would raise surface temperature above the boiling point for water.^[66] This was confirmed when Mariner 2 made radiometer measurements of the temperature in 1962. In 1967, Venera 4 confirmed that the atmosphere consisted primarily of carbon dioxide.^[66]

The upper atmosphere of Venus can be measured from Earth when the planet crosses the sun in a rare event known as a solar transit. The last solar transit of Venus occurred in 2012. Using quantitative astronomical spectroscopy, scientists were able to analyse sunlight that passed through the planet's atmosphere to reveal chemicals within it. As the technique to analyse light to discover information about a planet's

atmosphere only first showed results in 2001,^[67] this was the first opportunity to gain conclusive results in this way on the atmosphere of Venus since observation of solar transits began. This solar transit was a rare opportunity considering the lack of information on the atmosphere between 65 and 85 km.^[68] The solar transit in 2004 enabled astronomers to gather a large amount of data useful not only in determining the composition of the upper atmosphere of Venus, but also in refining techniques used in searching for extrasolar planets. The

atmosphere of mostly CO₂, absorbs near-infrared radiation, making it easy to observe. During the 2004 transit, the absorption in the atmosphere as a function of wavelength revealed the properties of the gases at that altitude. The Doppler shift of the gases also enabled wind patterns to be measured.^[69]

A solar transit of Venus is an extremely rare event, and the last solar transit of the planet before 2004 was in 1882. The most recent solar transit was in 2012; the next one will not occur until 2117.^{[68][69]}

Space missions

Recent and current spaceprobes

The Venus Express spacecraft formerly in orbit around the planet probed deeper into the atmosphere using infrared imaging spectroscopy in the 1–5 μm spectral range.^[4]

The JAXA probe Akatsuki, launched in May 2010, was intended to study the planet for a period of two years, including the structure and activity of the atmosphere, but it failed to enter Venus orbit in December 2010. A second attempt to achieve orbit succeeded 7 December 2015.^[70] One of its five cameras known as the "IR2" will be able to measure the atmosphere of the planet underneath its thick clouds, in addition to its movement and distribution of trace components. With a highly eccentric orbit (periapsis altitude of 400 km and apoapsis of 310,000 km), it will be able to take close-up photographs of the planet, and should also confirm the presence of both active volcanoes as well as lightning.^[71]

Approved missions in preparation

None known.

Proposed missions

The Venus In-Situ Explorer, proposed by NASA's New Frontiers program is a proposed probe which would aid in understanding the processes on the planet that led to climate change, as well as paving the way towards a later sample return mission.^[72]

Another craft called the Venus Mobile Explorer has been proposed by the Venus Exploration Analysis Group (VEXAG) to study the composition and isotopic measurements of the surface and the atmosphere, for about 90 days. A launch date has not yet been set.^[73]



Venus In-Situ Explorer proposed by NASA's New Frontiers program

After missions discovered the reality of the harsh nature of the planet's surface, attention shifted towards other targets such as Mars. There have been a number of proposed missions afterwards however, and many of these involve the little-known upper atmosphere. The Soviet Vega program in 1985 dropped two balloons into the atmosphere, but these were battery-powered and lasted for only about two Earth days each before running out of power. Since then, there has been no exploration of the upper atmosphere. In 2002, the NASA contractor Global Aerospace proposed a balloon that would be capable of staying in the upper atmosphere for hundreds of Earth days as opposed to two.^[74]

A solar flyer has also been proposed by Geoffrey A. Landis in place of a balloon,^[23] and the idea has been featured from time to time since the early 2000s. Venus has a high albedo, and reflects most of the sunlight that shines on it making the surface quite dark, the upper atmosphere at 60 km has an upward solar intensity of 90%, meaning that solar panels on both the top and the bottom of a craft could be used with nearly equal efficiency.^[38] In addition to this, the slightly lower gravity, high air pressure and slow rotation allowing for

perpetual solar power make this part of the planet ideal for exploration. The proposed flyer would operate best at an altitude where sunlight, air pressure, and wind speed would enable it to remain in the air perpetually, with slight dips down to lower altitudes for a few hours at a time before returning to higher altitudes. As sulfuric acid in the clouds at this height is not a threat for a properly shielded craft, this so-called "solar flyer" would be able to measure the area in between 45 km and 60 km indefinitely, for however long it takes for mechanical error or unforeseen problems to cause it to fail. Landis also proposed that rovers similar to Spirit and Opportunity could possibly explore the surface, with the difference being that Venus surface rovers would be "dumb" rovers controlled by radio signals from computers located in the flyer above,^[75] only requiring parts such as motors and transistors to withstand the surface conditions, but not weaker parts involved in microelectronics that could not be made resistant to the heat, pressure and acidic conditions.^[76]

Russian space plan for 2006–2015 involves a launch of Venera-D (Venus-D) probe around 2024.^[77] The main scientific goals of the Venera-D mission are investigation of the structure and chemical composition of the atmosphere and investigation of the upper atmosphere, ionosphere, electrical activity, magnetosphere and escape rate.^[78]

See also

- Ashen light, alleged atmospheric glow on Venus

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Atmospheric carbon cycle

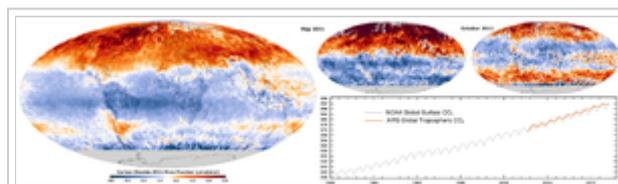
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The atmosphere is one of the Earth's major carbon reservoirs and an important component of the global carbon cycle, holding approximately 720 gigatons of carbon.^[1]

Atmospheric carbon plays an important role in the greenhouse effect. The most important carbon compound in this respect is the gas carbon dioxide (CO₂). Although it is

a small percentage of the atmosphere (approximately 0.04% on a molar basis), it plays a vital role in retaining

heat in the atmosphere and thus in the greenhouse effect.^[1] Other gases with effects on the climate containing carbon in the atmosphere are methane and chlorofluorocarbons (the latter is entirely anthropogenic). Emissions by humans in the past 200 years have almost doubled the amount carbon dioxide in the atmosphere.^{[1][2]}



2011 carbon dioxide mole fraction in the troposphere.

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Relevant gases

The concentration of mostly carbon-based greenhouse gases has increased dramatically since the onset of the industrial era. This makes an understanding of the carbon component of the atmosphere highly important. The two main carbon greenhouse gases are methane and carbon dioxide.^[3]

Methane

Methane (CH₄) is one of the more potent greenhouse gases and is mainly produced by the digestion or decay of biological organisms. It is considered the second most important greenhouse gas,^[3] yet the methane cycle in the atmosphere is currently only poorly understood.^[4] The amount of methane produced and absorbed yearly varies widely.^[3]

Large stores of methane can be found in the form of methane ice under permafrost and on continental shelves. Additional methane is produced by the anaerobic decay of organic material and is produced in organisms' digestive tracts, soil, etc. Natural methane production accounts 10-30% of global methane sources.^[5]

Anthropogenic methane is produced in various ways, e.g. by raising cattle or through the decay of trash in landfills. It is also produced by several industrial sources, including the mining and distribution of fossil fuels.^[4] More than 70% of atmospheric methane comes from biogenic sources. Methane levels have risen

plant matter decay at different rates - for example, woody substances retain their carbon longer than soft, leafy material. Active carbon in soils can stay sequestered for up to a thousand years, while inert carbon in soils can stay sequestered for more than a millennium.^[12]

Oceans

Large amounts of carbon are exchanged each year between the ocean and the atmosphere. A major controlling factor in oceanic-atmospheric carbon exchange is thermohaline circulation. In regions of ocean upwelling, carbon rich water from the deep ocean comes to the surface and releases carbon into the atmosphere as carbon dioxide. Large amounts of carbon dioxide are dissolved in cold water in higher latitudes. This water sinks down and brings the carbon into the deeper ocean levels, where it can stay for anywhere between decades and several centuries.^[1] Ocean circulation events cause this process to be variable. For example, during El Nino events there is less deep ocean upwelling, leading to lower outgassing of carbon dioxide into the atmosphere.^[11]

Biological processes also lead to ocean-atmosphere carbon exchange. Carbon dioxide equilibrates between the atmosphere and the ocean's surface layers. As autotrophs add or subtract carbon dioxide from the water through photosynthesis or respiration, they modify this balance, allowing the water to absorb more carbon dioxide or causing it to emit carbon dioxide into the atmosphere.^[1]

Geosphere

Carbon is generally exchanged very slowly between the atmosphere and geosphere. Two exceptions are volcanic eruptions and the combustion of fossil fuels, both of which release high amounts carbon into the atmosphere very quickly. Fresh silicate rock that is exposed through geological processes absorbs carbon from the atmosphere when it is exposed to air by the processes of weathering and erosion.

Anthropogenic sources

Human activities change the amount of carbon in the atmosphere directly through the burning of fossil fuels and other organic material, thus oxidizing the organic carbon and producing carbon dioxide.^{[13][14]} Another human-caused source of carbon dioxide is cement production. The burning of fossil fuels and cement production are the main reasons for the increase in atmospheric CO₂ since the beginning of the industrial era.^[3]

Other human-caused changes in the atmospheric carbon cycle are due to anthropogenic changes to carbon reservoirs. Deforestation, for example, decreases the biosphere's ability to absorb carbon, thus increasing the amount of carbon in the atmosphere.^[15]

As the industrial use of carbon by humans is a very new dynamic on a geologic scale, it is important to be able to track sources and sinks of carbon in the atmosphere. One way of doing so is by observing the proportion of stable carbon isotopes present in the atmosphere. The two main carbon isotopes are ¹²C and ¹³C. Plants absorb the lighter isotope, ¹²C, more readily than ¹³C.^[16] Because fossil fuels originate mainly from plant matter, the ¹³C/¹²C ratio in the atmosphere falls when large amounts of fossil fuels are burned, releasing ¹²C. Conversely, an increase in the ¹³C/¹²C in the atmosphere suggests a higher biospheric carbon uptake.^[12] The ratio of the annual increase in atmospheric CO₂ compared to CO₂ emissions from fossil fuel and cement manufactured is called the "airborne fraction."^[17] The airborne fraction has been around 60% since the 1950s, indicating that about 60% of the new carbon dioxide in the atmosphere each year originated from human sources.^[3] For clarity, this is not meant to suggest that 60% of the uptake of carbon dioxide into the atmosphere comes from human activity. It means that the atmosphere exchanges around 210 giga-tonnes of carbon annually, but absorbs between 6 and 10 giga-tonnes more than it loses. Of this net gain, about 60% is attributable to the burning of fossil fuels.

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External links

- Advanced Global Atmospheric Gases Experiment
- Carbon Dioxide - Global Circulation (NASA; 13 December 2016)
- Carbon Dioxide - Information Analysis Center
- World Data Centre for Greenhouse Gases

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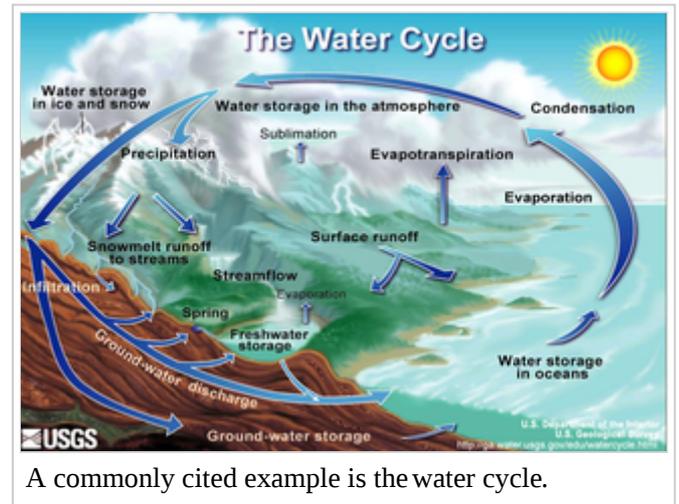
Categories: Carbon | Atmospheric chemistry

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Biogeochemical cycle

From Wikipedia, the free encyclopedia

In Earth science, a **biogeochemical cycle** or **substance turnover** or **cycling of substances** is a pathway by which a chemical substance moves through both the biotic (biosphere) and abiotic (lithosphere, atmosphere, and hydrosphere) components of Earth. A cycle is a series of change which comes back to the starting point and which can be repeated.^{[1][2]} Water, for example, is always recycled through the water cycle, as shown in the diagram. The water undergoes evaporation, condensation, and precipitation, falling back to Earth. Elements, chemical compounds, and other forms of matter are passed from one organism to another and from one part of the biosphere to another through biogeochemical cycles.^{[1][2]}



A commonly cited example is the water cycle.

The term "biogeochemical" tells us that biological, geological and chemical factors are all involved. The circulation of chemical nutrients like carbon, oxygen, nitrogen, phosphorus, calcium, and water etc. through the biological and physical world are known as "biogeochemical cycles". In effect, the element is recycled, although in some cycles there may be places (called *reservoirs*) where the element is accumulated or held for a long period of time (such as an ocean or lake for water).^{[1][2]}

Important cycles

The most well-known and important biogeochemical cycles, for example,

- the carbon cycle,
- the nitrogen cycle,
- the oxygen cycle,
- the phosphorus cycle,
- the sulfur cycle,
- the water cycle,
- and the rock cycle.

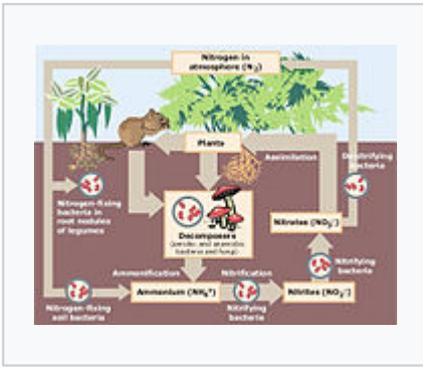
There are many biogeochemical cycles that are currently being studied for the first time as climate change and human impacts are drastically changing the speed, intensity, and balance of these relatively unknown cycles. These newly studied biogeochemical cycles include

- the mercury cycle,^[3] and
- the human-caused cycle of atrazine, which may affect certain species.

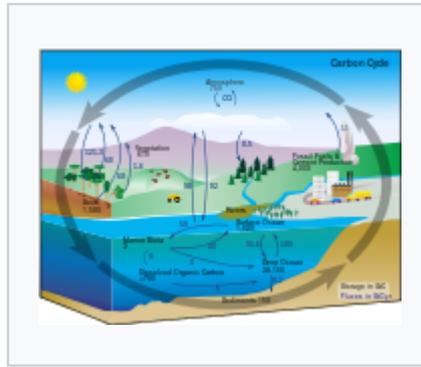
Biogeochemical cycles always involve hot equilibrium states: a balance in the cycling of the element between compartments. However, overall balance may involve compartments distributed on a global scale.

As biogeochemical cycles describe the movements of substances on the entire globe, the study of these is inherently multidisciplinary. The carbon cycle may be related to research in ecology and atmospheric sciences.^[4] Biochemical dynamics would also be related to the fields of geology and pedology (soil study).^[5]

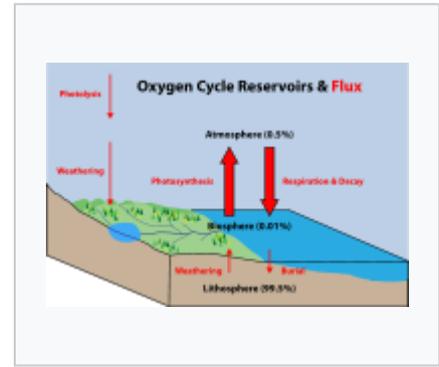
Global biogeochemical cycles critical for life



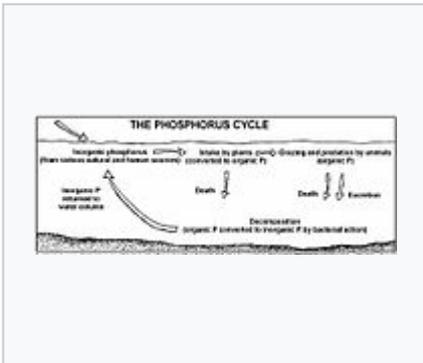
Nitrogen cycle



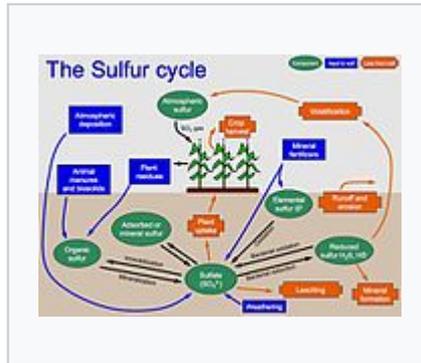
Carbon cycle



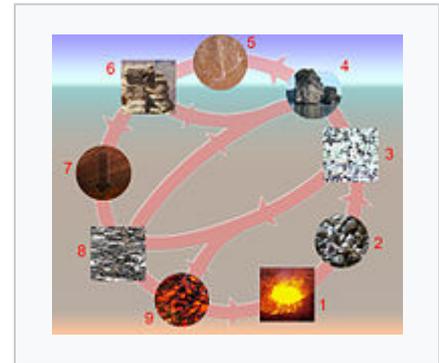
Oxygen cycle



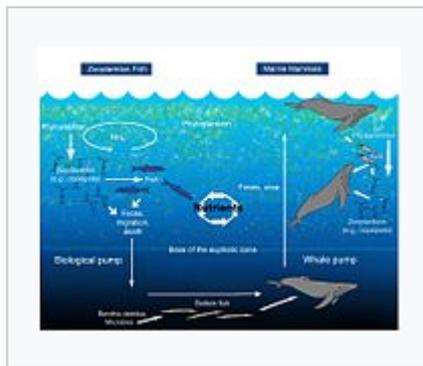
Phosphorus cycle



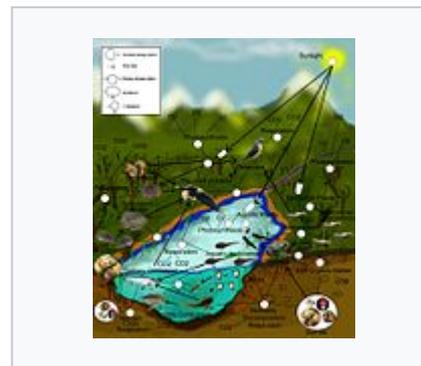
Sulfur cycle



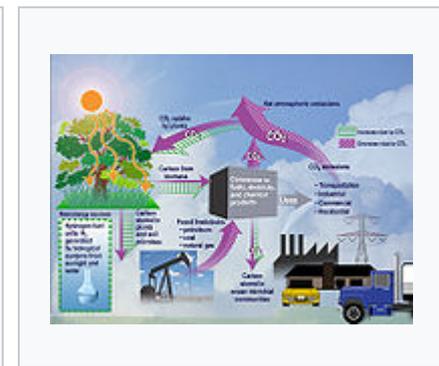
Rock cycle



Nutrient cycle



Food cycle



Carbon dioxide

See also

- Biodiversity
- Bioregion
- Earth science
- Ecology
- Ecosystem
- Soil zoology
- GEOTRACES
- Holistic management
- Natural environment
- Nature
- Nutrient cycle
- Planetary boundaries

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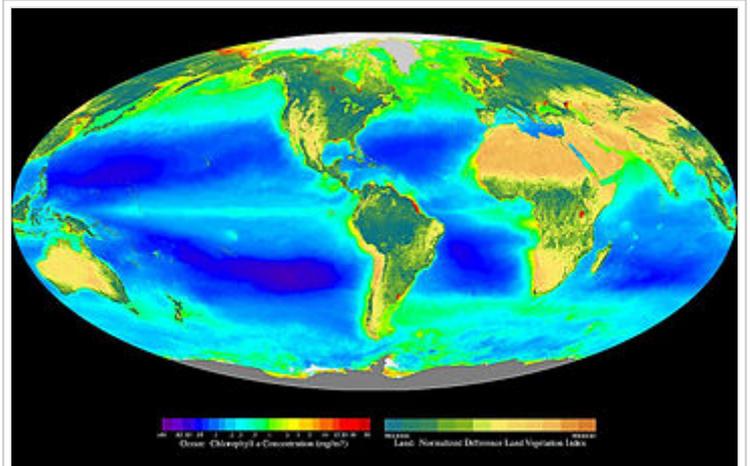
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Biosphere

From Wikipedia, the free encyclopedia

The **biosphere** (from Greek βίος *bíos* "life" and σφαῖρα *sphaira* "sphere") also known as the **ecosphere** (from Greek οἶκος *oîkos* "environment" and σφαῖρα), is the worldwide sum of all ecosystems. The two joined words are "bio" and "sphere". It can also be termed as the zone of life on Earth, a closed system (apart from solar and cosmic radiation and heat from the interior of the Earth), and largely self-regulating.^[1] By the most general biophysiological definition, the biosphere is the global ecological system integrating all living beings and their relationships, including their interaction with the elements of the lithosphere, geosphere, hydrosphere, and atmosphere. The biosphere is postulated to have evolved, beginning with a process of biopoiesis (life created naturally from non-living matter, such as simple organic



A false-color composite of global oceanic and terrestrial photoautotroph abundance, from September 2001 to August 2017. Provided by the SeaWiFS Project, NASA/Goddard Space Flight Center and ORBIMAGE.

compounds) or biogenesis (life created from living matter), at least some 3.5 billion years ago.^{[2][3]} The earliest evidence for life on Earth includes biogenic graphite found in 3.7 billion-year-old metasedimentary rocks from Western Greenland^[4] and microbial mat fossils found in 3.48 billion-year-old sandstone from Western Australia.^{[5][6]} More recently, in 2015, "remains of biotic life" were found in 4.1 billion-year-old rocks in Western Australia.^{[7][8]} In 2017, putative fossilized microorganisms (or microfossils) were announced to have been discovered in hydrothermal vent precipitates in the Nuvvuagittuq Belt of Quebec, Canada that were as old as 4.28 billion years, the oldest record of life on earth, suggesting "an almost instantaneous emergence of life" after ocean formation 4.4 billion years ago, and not long after the formation of the Earth 4.54 billion years ago.^{[9][10][11][12]} According to one of the researchers, "If life arose relatively quickly on Earth ... then it could be common in the universe."^[7]

In a general sense, biospheres are any closed, self-regulating systems containing ecosystems. This includes artificial biospheres such as Biosphere 2 and BIOS-3, and potentially ones on other planets or moons.^[13]

Contents

- 1 Origin and use of the term
 - 1.1 Narrow definition
- 2 Extent of Earth's biosphere
- 3 Specific biospheres
- 4 Extraterrestrial biospheres
- 5 See also
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Origin and use of the term

The term "biosphere" was coined by geologist Eduard Suess in 1875, which he defined as the place on Earth's surface where life dwells.^[14]

While the concept has a geological origin, it is an indication of the effect of both Charles Darwin and Matthew F. Maury on the Earth sciences. The biosphere's ecological context comes from the 1920s (see Vladimir I. Vernadsky), preceding the 1935 introduction of the term "ecosystem" by Sir Arthur Tansley (see ecology history). Vernadsky defined ecology as the science of the biosphere. It is an interdisciplinary concept for integrating astronomy, geophysics, meteorology, biogeography, evolution, geology, geochemistry, hydrology and, generally speaking, all life and Earth sciences.

Narrow definition

Geochemists define the biosphere as being the total sum of living organisms (the "biomass" or "biota" as referred to by biologists and ecologists). In this sense, the biosphere is but one of four separate components of the geochemical model, the other three being *geosphere*, *hydrosphere*, and *atmosphere*. When these four component spheres are combined into one system, it is known as the Ecosphere. This term was coined during the 1960s and encompasses both biological and physical components of the planet.^[15]

The Second International Conference on Closed Life Systems defined *biospherics* as the science and technology of analogs and models of Earth's biosphere; i.e., artificial Earth-like biospheres.^[16] Others may include the creation of artificial non-Earth biospheres—for example, human-centered biospheres or a native Martian biosphere—as part of the topic of biospherics.

Extent of Earth's biosphere



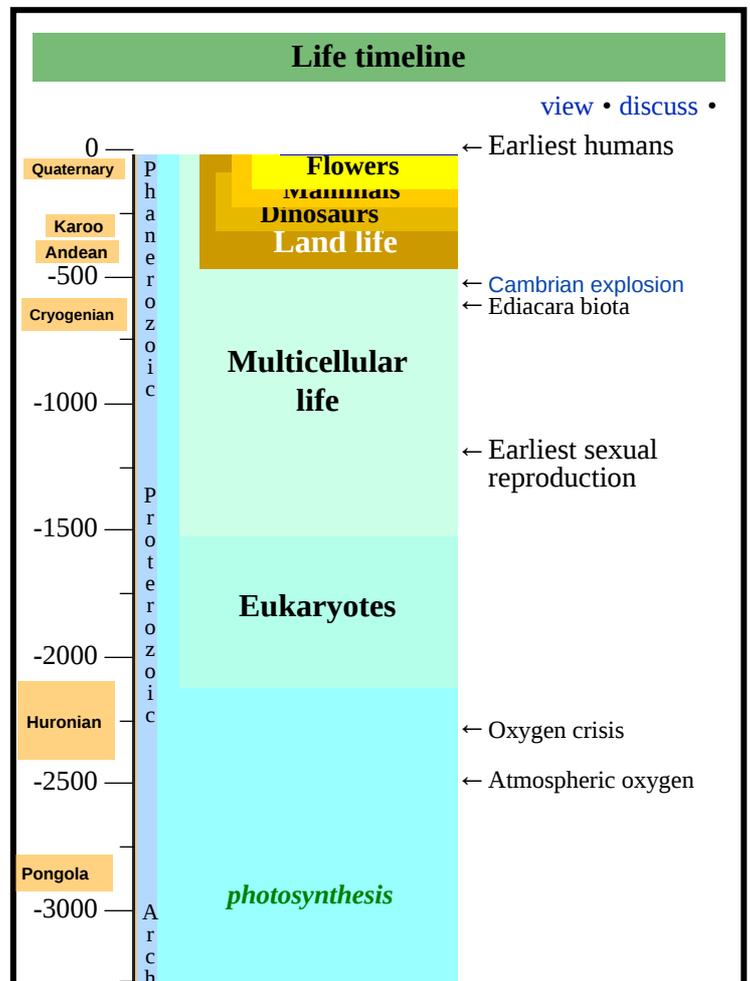
Water covers 71% of the Earth's surface. Image is the Blue Marble photographed from Apollo 17.

Every part of the planet, from the polar ice caps to the equator, features life of some kind. Recent advances in

microbiology have demonstrated that microbes live deep beneath the Earth's terrestrial surface, and that the total mass of microbial life in so-called "uninhabitable zones" may, in biomass, exceed all animal and plant life on the surface. The actual thickness of the biosphere on earth is difficult to measure. Birds typically fly at altitudes

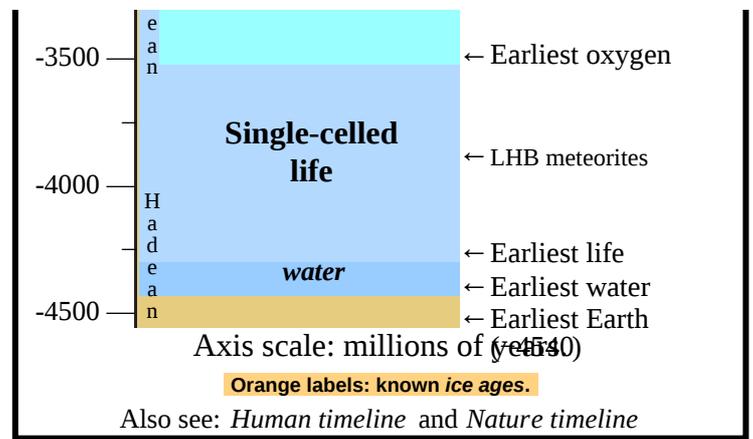


A beach scene on Earth, simultaneously showing the lithosphere (ground), hydrosphere (ocean) and atmosphere (air)



as high as 1,800 m (5,900 ft; 1.1 mi) and fish live as much as 8,372 m (27,467 ft; 5.202 mi) underwater in the Puerto Rico Trench.^[2]

There are more extreme examples for life on the planet: Rüppell's vulture has been found at altitudes of 11,300 m (37,100 ft; 7.0 mi); bar-headed geese migrate at altitudes of at least 8,300 m (27,200 ft; 5.2 mi); yaks live at elevations as high as 5,400 m (17,700 ft; 3.4 mi) above sea level; mountain goats live up to 3,050 m (10,010 ft; 1.90 mi). Herbivorous animals at these elevations depend on lichens, grasses, and herbs.



Life forms live in every part of the Earth's biosphere, including soil, hot springs, inside rocks at least 19 km (12 mi) deep underground, the deepest parts of the ocean, and at least 64 km (40 mi) high in the atmosphere.^{[17][18][19]} Microorganisms, under certain test conditions, have been observed to thrive in the vacuum of outer space.^{[20][21]} The total amount of soil and subsurface bacterial carbon is estimated as 5×10^{17} g, or the "weight of the United Kingdom".^[17] The mass of prokaryote microorganisms—which includes bacteria and archaea, but not the nucleated eukaryote microorganisms—may be as much as 0.8 trillion tons of carbon (of the total biosphere mass, estimated at between 1 and 4 trillion tons).^[22] Barophilic marine microbes have been found at more than a depth of 10,000 m (33,000 ft; 6.2 mi) in the Mariana Trench, the deepest spot in the Earth's oceans.^[23] In fact, single-celled life forms have been found in the deepest part of the Mariana Trench, by the Challenger Deep, at depths of 11,034 m (36,201 ft; 6.856 mi).^{[24][25][26]} Other researchers reported related studies that microorganisms thrive inside rocks up to 580 m (1,900 ft; 0.36 mi) below the sea floor under 2,590 m (8,500 ft; 1.61 mi) of ocean off the coast of the northwestern United States,^{[25][27]} as well as 2,400 m (7,900 ft; 1.5 mi) beneath the seabed off Japan.^[28] Culturable thermophilic microbes have been extracted from cores drilled more than 5,000 m (16,000 ft; 3.1 mi) into the Earth's crust in Sweden,^[29] from rocks between 65–75 °C (149–167 °F). Temperature increases with increasing depth into the Earth's crust. The rate at which the temperature increases depends on many factors, including type of crust (continental vs. oceanic), rock type, geographic location, etc. The greatest known temperature at which microbial life can exist is 122 °C (252 °F) (*Methanopyrus kandleri* Strain 116), and it is likely that the limit of life in the "deep biosphere" is defined by temperature rather than absolute depth. On 20 August 2014, scientists confirmed the existence of microorganisms living 800 m (2,600 ft; 0.50 mi) below the ice of Antarctica.^{[30][31]} According to one researcher, "You can find microbes everywhere — they're extremely adaptable to conditions, and survive wherever they are."^[25]

Our biosphere is divided into a number of biomes, inhabited by fairly similar flora and fauna. On land, biomes are separated primarily by latitude. Terrestrial biomes lying within the Arctic and Antarctic Circles are relatively barren of plant and animal life, while most of the more populous biomes lie near the equator.

Specific biospheres

For this list, if a word is followed by a number, it is usually referring to a specific system or number. Thus:

- Biosphere 1, the planet Earth.
- Biosphere 2, laboratory in Arizona, United States, which contains 3.15 acres (13,000 m²) of closed ecosystem.
- BIOS-3, a closed ecosystem at the Institute of Biophysics in Krasnoyarsk, Siberia, in what was then the Soviet Union.
- Biosphere J (CEEF, Closed Ecology Experiment Facilities), an experiment in Japan.^{[32][33]}

Extraterrestrial biospheres

No biospheres have been detected beyond the Earth; therefore, the existence of extraterrestrial biospheres remains hypothetical. The rare Earth hypothesis suggests they should be very rare, save ones composed of microbial life only.^[34] On the other hand, Earth analogs may be quite numerous, at least in the Milky Way galaxy.^[35] Three of the planets discovered orbiting TRAPPIST-1 could possibly contain biospheres.^[36] Given limited understanding of abiogenesis, it is currently unknown what percentage of these planets actually develop biospheres.

It is also possible that artificial biospheres will be created during the future, for example on Mars.^[37] The process of creating an uncontained system that mimics the function of Earth's biosphere is called terraforming.^[38]

See also

- Abiogenesis
- Biosphere reserve
- Closed ecological system
- Cryosphere
- Ecosystem
- Geosphere
- Habitable zone
- Homeostasis
- Life support system
- Montreal Biosphère
- Noogenesis
- Noosphere
- Shadow biosphere
- Simple biosphere model
- Soil biomantle
- Thomas Gold
- Wardian case
- Winogradsky column

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Further reading

- *The Biosphere* (A Scientific American Book), San Francisco, W.H. Freeman and Co., 1970, ISBN 0-7167-0945-7. This book, originally a 1970 *Scientific American* issue, covers virtually every major concern and concept since debated regarding materials and energy resources, population trends, and environmental degradation.

External links

- Biosphere Definition
- Article on the Biosphere at Encyclopedia of Earth
- GLOBIO.info, an ongoing programme to map the past, current and future impacts of human activities on the biosphere
- Paul Crutzen Interview, freeview video of Paul Crutzen Nobel Laureate for his work on decomposition of ozone talking to Harry Kroto Nobel Laureate by the Vega Science Trust.
- Atlas of the Biosphere

	Look up biosphere in Wiktionary, the free dictionary.
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Categories: Oceanography | Superorganisms | Biological systems

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Black-body radiation

From Wikipedia, the free encyclopedia

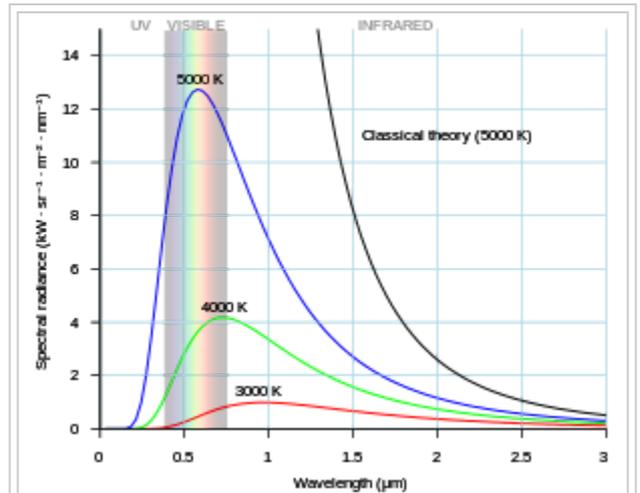
Black-body radiation is the thermal electromagnetic radiation within or surrounding a body in thermodynamic equilibrium with its environment, or emitted by a black body (an opaque and non-reflective body). It has a specific spectrum and intensity that depends only on the body's temperature, which is assumed for the sake of calculations and theory to be uniform and constant.^{[1][2][3][4]}

The thermal radiation spontaneously emitted by many ordinary objects can be approximated as black-body radiation. A perfectly insulated enclosure that is in thermal equilibrium internally contains black-body radiation and will emit it through a hole made in its wall, provided the hole is small enough to have negligible effect upon the equilibrium.

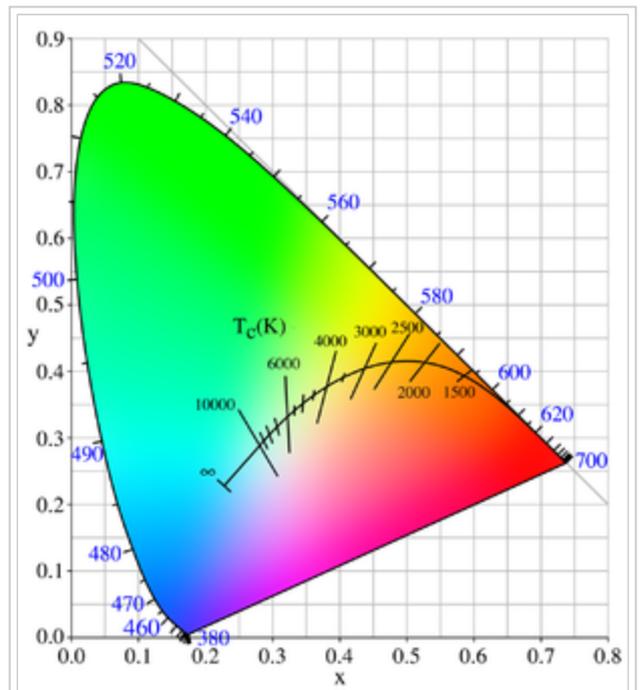
A black-body at room temperature appears black, as most of the energy it radiates is infra-red and cannot be perceived by the human eye. Because the human eye cannot perceive color at very low light intensities, a black body, viewed in the dark at the lowest just faintly visible temperature, subjectively appears grey (but only because the human eye is sensitive only to black and white at very low intensities - in reality, the frequency of the light in the visible range would still be red, although the intensity would be too low to discern as red), even though its objective physical spectrum peaks in the infrared range.^[5] When it becomes a little hotter, it appears dull red. As its temperature increases further it eventually becomes blue-white.

Although planets and stars are neither in thermal equilibrium with their surroundings nor perfect black bodies, black-body radiation is used as a first approximation for the energy they emit.^[6] Black holes are near-perfect black bodies, in the sense that they absorb all the radiation that falls on them. It has been proposed that they emit black-body radiation (called Hawking radiation), with a temperature that depends on the mass of the black hole.^[7]

The term *black body* was introduced by Gustav Kirchhoff in 1860. Black-body radiation is also called thermal radiation, *cavity radiation*, *complete radiation* or *temperature radiation*.



As the temperature decreases, the peak of the black-body radiation curve moves to lower intensities and longer wavelengths. The black-body radiation graph is also compared with the classical model of Rayleigh and Jeans.



The color (chromaticity) of black-body radiation depends on the temperature of the black body; the locus of such colors, shown here in CIE 1931 x,y space, is known as the Planckian locus.

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Spectrum

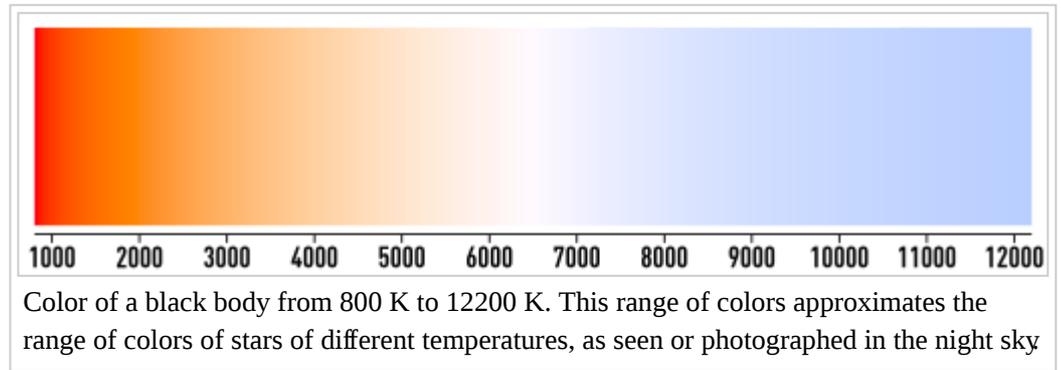
Black-body radiation has a characteristic, continuous frequency spectrum that depends only on the body's temperature,^[8] called the Planck spectrum or Planck's law. The spectrum is peaked at a characteristic frequency that shifts to higher frequencies with increasing temperature, and at room temperature most of the emission is in the infrared region of the electromagnetic spectrum.^{[9][10][11]} As the temperature increases past about 500 degrees Celsius, black bodies start to emit significant amounts of visible light. Viewed in the dark by the human eye, the first faint glow appears as a "ghostly" grey (the visible light is actually red, but low intensity light activates only the eye's grey-level sensors). With rising temperature, the glow becomes visible even when there is some background surrounding light: first as a dull red, then yellow, and eventually a "dazzling bluish-white" as the temperature rises.^{[12][13]} When the body appears white, it is emitting a substantial fraction of its energy as ultraviolet radiation. The Sun, with an effective temperature of approximately 5800 K,^[14] is an approximate black body with an emission spectrum peaked in the central, yellow-green part of the visible spectrum, but with significant power in the ultraviolet as well.

Black-body radiation provides insight into the thermodynamic equilibrium state of cavity radiation. If each Fourier mode of the equilibrium radiation in an otherwise empty cavity with perfectly reflective walls is considered as a degree of freedom capable of exchanging energy, then, according to the equipartition theorem of classical physics, there would be an equal amount of energy in each mode. Since there are an infinite number of modes this implies infinite heat capacity (infinite energy at any non-zero temperature), as well as an unphysical spectrum of emitted radiation that grows without bound with increasing frequency, a problem known as the ultraviolet catastrophe. Instead, in quantum theory the occupation numbers of the modes are quantized, cutting off the spectrum at high frequency in agreement with experimental observation and resolving the catastrophe. The study of the laws of black bodies and the failure of classical physics to describe them helped establish the foundations of quantum mechanics.

Explanation

All normal (baryonic) matter emits electromagnetic radiation when it has a temperature above absolute zero. The radiation represents a conversion of a body's thermal energy into electromagnetic energy, and is therefore called thermal radiation. It is a spontaneous process of radiative distribution of entropy.

Conversely all normal matter absorbs electromagnetic radiation to some degree. An object that absorbs all radiation falling on it, at all wavelengths, is called a black body. When a black body is at a uniform temperature, its emission has a characteristic frequency distribution that depends on the temperature. Its emission is called black-body radiation.



The concept of the black body is an idealization, as perfect black bodies do not exist in nature.^[15] Graphite and lamp black, with emissivities greater than 0.95, however, are good approximations to a black material. Experimentally, black-body radiation may be established best as the ultimately stable steady state equilibrium radiation in a cavity in a rigid body, at a uniform temperature, that is entirely opaque and is only partly reflective.^[15] A closed box of graphite walls at a constant temperature with a small hole on one side produces a good approximation to ideal black-body radiation emanating from the opening.^{[16][17]}

Black-body radiation has the unique absolutely stable distribution of radiative intensity that can persist in thermodynamic equilibrium in a cavity.^[15] In equilibrium, for each frequency the total intensity of radiation that is emitted and reflected from a body (that is, the net amount of radiation leaving its surface, called the *spectral radiance*) is determined solely by the equilibrium temperature, and does not depend upon the shape, material or structure of the body.^[18] For a black body (a perfect absorber) there is no reflected radiation, and so the spectral radiance is due entirely to emission. In addition, a black body is a diffuse emitter (its emission is independent of direction). Consequently, black-body radiation may be viewed as the radiation from a black body at thermal equilibrium.

Black-body radiation becomes a visible glow of light if the temperature of the object is high enough. The Draper point is the temperature at which all solids glow a dim red, about 798 K.^{[19][20]} At 1000 K, a small opening in the wall of a large uniformly heated opaque-walled cavity (let us call it an oven), viewed from outside, looks red; at 6000 K, it looks white. No matter how the oven is constructed, or of what material, as long as it is built so that almost all light entering is absorbed by its walls, it will contain a good approximation to black-body radiation. The spectrum, and therefore color, of the light that comes out will be a function of the cavity temperature alone. A graph of the amount of energy inside the oven per unit volume and per unit frequency interval plotted versus frequency, is called the *black-body curve*. Different curves are obtained by varying the temperature.

Two bodies that are at the same temperature stay in mutual thermal equilibrium, so a body at temperature T surrounded by a cloud of light at temperature T on average will emit as much light into the cloud as it absorbs, following Prevost's exchange principle, which refers to radiative equilibrium. The principle of detailed balance says that in thermodynamic equilibrium every elementary process works equally in its forward and backward sense.^{[21][22]} Prevost also showed that the emission from a body is logically determined solely by its own internal state. The causal effect of thermodynamic absorption on thermodynamic (spontaneous) emission is not direct, but is only indirect as it affects the internal state of the body. This means that at thermodynamic equilibrium the amount of every wavelength in every direction of thermal radiation emitted by a body at temperature T , black or not, is equal to the corresponding amount that the body absorbs because it is surrounded by light at temperature T .^[23]

When the body is black, the absorption is obvious: the amount of light absorbed is all the light that hits the surface. For a black body much bigger than the wavelength, the light energy absorbed at any wavelength λ per unit time is strictly proportional to the black-body curve. This means that the black-body curve is the amount of light energy emitted by a black body, which justifies the name. This is the condition for the applicability of



The temperature of a Pāhoehoe lava flow can be estimated by observing its color. The result agrees well with other measurements of temperatures of lava flows at about 1,000 to 1,200 °C (1,830 to 2,190 °F).

Kirchhoff's law of thermal radiation: the black-body curve is characteristic of thermal light, which depends only on the temperature of the walls of the cavity, provided that the walls of the cavity are completely opaque and are not very reflective, and that the cavity is in thermodynamic equilibrium.^[24] When the black body is small, so that its size is comparable to the wavelength of light, the absorption is modified, because a small object is not an efficient absorber of light of long wavelength, but the principle of strict equality of emission and absorption is always upheld in a condition of thermodynamic equilibrium.

In the laboratory, black-body radiation is approximated by the radiation from a small hole in a large cavity, a hohlraum, in an entirely opaque body that is only partly reflective, that is maintained at a constant temperature. (This technique leads to the alternative term *cavity radiation*.) Any light entering the hole would have to reflect off the walls of the cavity multiple times before it

escaped, in which process it is nearly certain to be absorbed. Absorption occurs regardless of the wavelength of the radiation entering (as long as it is small compared to the hole). The hole, then, is a close approximation of a theoretical black body and, if the cavity is heated, the spectrum of the hole's radiation (i.e., the amount of light emitted from the hole at each wavelength) will be continuous, and will depend only on the temperature and the fact that the walls are opaque and at least partly absorptive, but not on the particular material of which they are built nor on the material in the cavity (compare with emission spectrum).

Calculating the black-body curve was a major challenge in theoretical physics during the late nineteenth century. The problem was solved in 1901 by Max Planck in the formalism now known as Planck's law of black-body radiation.^[25] By making changes to Wien's radiation law (not to be confused with Wien's displacement law) consistent with thermodynamics and electromagnetism, he found a mathematical expression fitting the experimental data satisfactorily. Planck had to assume that the energy of the oscillators in the cavity was quantized, i.e., it existed in integer multiples of some quantity. Einstein built on this idea and proposed the quantization of electromagnetic radiation itself in 1905 to explain the photoelectric effect. These theoretical advances eventually resulted in the superseding of classical electromagnetism by quantum electrodynamics. These quanta were called photons and the black-body cavity was thought of as containing a gas of photons. In addition, it led to the development of quantum probability distributions, called Fermi–Dirac statistics and Bose–Einstein statistics, each applicable to a different class of particles, fermions and bosons.

The wavelength at which the radiation is strongest is given by Wien's displacement law, and the overall power emitted per unit area is given by the Stefan–Boltzmann law. So, as temperature increases, the glow color changes from red to yellow to white to blue. Even as the peak wavelength moves into the ultra-violet, enough radiation continues to be emitted in the blue wavelengths that the body will continue to appear blue. It will never become invisible—indeed, the radiation of visible light increases monotonically with temperature.^[26]

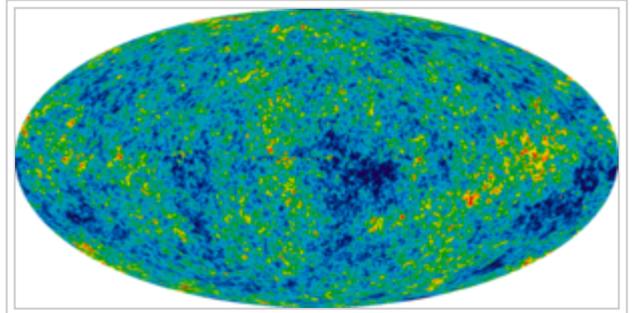
The radiance or observed intensity is not a function of direction. Therefore, a black body is a perfect Lambertian radiator.

Real objects never behave as full-ideal black bodies, and instead the emitted radiation at a given frequency is a fraction of what the ideal emission would be. The emissivity of a material specifies how well a real body radiates energy as compared with a black body. This emissivity depends on factors such as temperature, emission angle, and wavelength. However, it is typical in engineering to assume that a surface's spectral emissivity and absorptivity do not depend on wavelength, so that the emissivity is a constant. This is known as the *gray body* assumption.

With non-black surfaces, the deviations from ideal black-body behavior are determined by both the surface structure, such as roughness or granularity, and the chemical composition. On a "per wavelength" basis, real objects in states of local thermodynamic equilibrium still follow Kirchhoff's Law: emissivity equals

absorptivity, so that an object that does not absorb all incident light will also emit less radiation than an ideal black body; the incomplete absorption can be due to some of the incident light being transmitted through the body or to some of it being reflected at the surface of the body.

In astronomy, objects such as stars are frequently regarded as black bodies, though this is often a poor approximation. An almost perfect black-body spectrum is exhibited by the cosmic microwave background radiation. Hawking radiation is the hypothetical black-body radiation emitted by black holes, at a temperature that depends on the mass, charge, and spin of the hole. If this prediction is correct, black holes will very gradually shrink and evaporate over time as they lose mass by the emission of photons and other particles.



9-year WMAP image (2012) of the cosmic microwave background radiation across the universe.^{[27][28]}

A black body radiates energy at all frequencies, but its intensity rapidly tends to zero at high frequencies (short wavelengths). For example, a black body at room temperature (300 K) with one square meter of surface area will emit a photon in the visible range (390–750 nm) at an average rate of one photon every 41 seconds, meaning that for most practical purposes, such a black body does not emit in the visible range.^[29]

Equations

Planck's law of black-body radiation

Planck's law states that^[30]

$$B_{\nu}(T) = \frac{2h\nu^3}{c^2} \frac{1}{e^{\frac{h\nu}{kT}} - 1},$$

where

- $B_{\nu}(T)$ is the spectral radiance (the power per unit solid angle and per unit of area normal to the propagation) density of frequency ν radiation per unit frequency at thermal equilibrium at temperature T .
- h is the Planck constant;
- c is the speed of light in a vacuum;
- k is the Boltzmann constant;
- ν is the frequency of the electromagnetic radiation;
- T is the absolute temperature of the body.

For a black body surface the spectral radiance density (defined per unit of area normal to the propagation) is independent of the angle θ of emission with respect to the normal. However, this means that, following Lambert's cosine law, $B_{\nu}(T) \cos \theta$ is the radiance density per unit area of emitting surface as the surface area involved in generating the radiance is increased by a factor $1/\cos \theta$ with respect to an area normal to the propagation direction. At oblique angles, the solid angle spans involved do get smaller, resulting in lower aggregate intensities.

Wien's displacement law

Wien's displacement law shows how the spectrum of black-body radiation at any temperature is related to the spectrum at any other temperature. If we know the shape of the spectrum at one temperature, we can calculate the shape at any other temperature. Spectral intensity can be expressed as a function of wavelength or of frequency.

A consequence of Wien's displacement law is that the wavelength at which the intensity *per unit wavelength* of the radiation produced by a black body is at a maximum, λ_{\max} , is a function only of the temperature:

$$\lambda_{\max} = \frac{b}{T},$$

where the constant b , known as Wien's displacement constant, is equal to $2.897\,7729(17) \times 10^{-3} \text{ K m}$.^[31]

Planck's law was also stated above as a function of frequency. The intensity maximum for this is given by

$$\nu_{\max} = T \times 58.8 \text{ GHz/K.}^{[32]}$$

Stefan–Boltzmann law

By integrating $B_\nu(T)$ over the frequency the integrated radiance L is

$$L = \frac{2\pi^5}{15} \frac{k^4 T^4}{c^2 h^3} \frac{1}{\pi} =: \sigma T^4 \frac{1}{\pi}$$

by using $\int_0^\infty dx \frac{x^3}{e^x - 1} = \frac{\pi^4}{15}$ with $x \equiv \frac{h\nu}{kT}$ and with $\sigma \equiv \frac{2\pi^5}{15} \frac{k^4}{c^2 h^3} = 5.670373 \times 10^{-8} \frac{W}{m^2 K^4}$ being the Stefan–Boltzmann constant. The radiance L is then

$$\sigma T^4 \frac{\cos \theta}{\pi}$$

per unit of emitting surface.

On a side note, at a distance d , the intensity dI per area dA of radiating surface is the useful expression

$$dI = \sigma T^4 \frac{\cos \theta}{\pi d^2} dA$$

when the receiving surface is perpendicular to the radiation.

By subsequently integrating over the solid angle Ω (where $\theta < \pi/2$) the Stefan–Boltzmann law is calculated, stating that the power j^* emitted per unit area of the surface of a black body is directly proportional to the fourth power of its absolute temperature:

$$j^* = \sigma T^4,$$

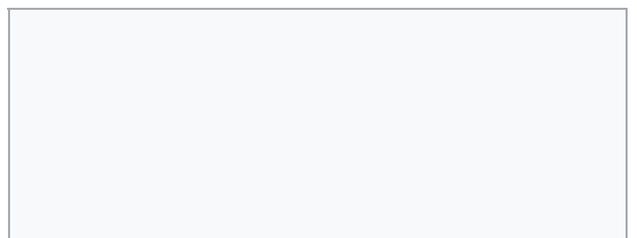
by using

$$\int \cos \theta d\Omega = \int_0^{2\pi} \int_0^{\pi/2} \cos \theta \sin \theta d\theta d\phi = \pi.$$

Human-body emission

The human body radiates energy as infrared light. The net power radiated is the difference between the power emitted and the power absorbed:

$$P_{\text{net}} = P_{\text{emit}} - P_{\text{absorb}}.$$



Applying the Stefan–Boltzmann law,

$$P_{\text{net}} = A\sigma\epsilon (T^4 - T_0^4),$$

where A and T are the body surface area and temperature, ϵ is the emissivity, and T_0 is the ambient temperature.

The total surface area of an adult is about 2 m^2 , and the mid- and far-infrared emissivity of skin and most clothing is near unity, as it is for most nonmetallic surfaces.^{[33][34]} Skin temperature is about $33 \text{ }^\circ\text{C}$,^[35] but clothing reduces the surface temperature to about $28 \text{ }^\circ\text{C}$ when the ambient temperature is $20 \text{ }^\circ\text{C}$.^[36] Hence, the net radiative heat loss is about

$$P_{\text{net}} = 100 \text{ W}.$$

The total energy radiated in one day is about 9 MJ , or 2000 kcal (food calories). Basal metabolic rate for a 40-year-old male is about $35 \text{ kcal}/(\text{m}^2 \cdot \text{h})$,^[37] which is equivalent to 1700 kcal per day, assuming the same 2 m^2 area. However, the mean metabolic rate of sedentary adults is about 50% to 70% greater than their basal rate.^[38]

There are other important thermal-loss mechanisms, including convection and evaporation. Conduction is negligible – the Nusselt number is much greater than unity. Evaporation by perspiration is only required if radiation and convection are insufficient to maintain a steady-state temperature (but evaporation from the lungs occurs regardless). Free-convection rates are comparable, albeit somewhat lower, than radiative rates.^[39] Thus, radiation accounts for about two-thirds of thermal energy loss in cool, still air. Given the approximate nature of many of the assumptions, this can only be taken as a crude estimate. Ambient air motion, causing forced convection, or evaporation reduces the relative importance of radiation as a thermal-loss mechanism.

Application of Wien's law to human-body emission results in a peak wavelength of

$$\lambda_{\text{peak}} = \frac{2.898 \times 10^{-3} \text{ K} \cdot \text{m}}{305 \text{ K}} = 9.50 \text{ } \mu\text{m}.$$

For this reason, thermal imaging devices for human subjects are most sensitive in the 7–14 micrometer range.

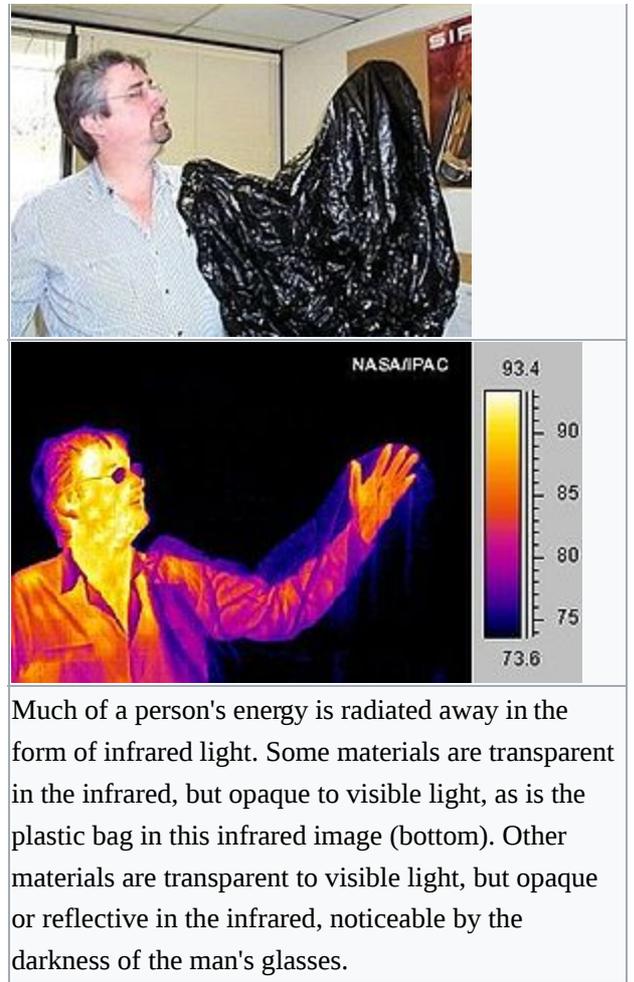
Temperature relation between a planet and its star

The black-body law may be used to estimate the temperature of a planet orbiting the Sun.

The temperature of a planet depends on several factors:

- Incident radiation from its star
- Emitted radiation of the planet, e.g., Earth's infrared glow
- The albedo effect causing a fraction of light to be reflected by the planet
- The greenhouse effect for planets with an atmosphere
- Energy generated internally by a planet itself due to radioactive decay, tidal heating, and adiabatic contraction due by cooling.

The analysis only considers the Sun's heat for a planet in a Solar System.



The Stefan–Boltzmann law gives the total power (energy/second) the Sun is emitting:

$$P_{S \text{ emt}} = 4\pi R_S^2 \sigma T_S^4 \quad (1)$$

where

σ is the Stefan–Boltzmann constant,
 T_S is the effective temperature of the Sun, and
 R_S is the radius of the Sun.

The Sun emits that power equally in all directions. Because of this, the planet is hit with only a tiny fraction of it. The power from the Sun that strikes the planet (at the top of the atmosphere) is:

$$P_{SE} = P_{S \text{ emt}} \left(\frac{\pi R_E^2}{4\pi D^2} \right) \quad (2)$$

where

R_E is the radius of the planet and
 D is the distance between the Sun and the planet.

Because of its high temperature, the Sun emits to a large extent in the ultraviolet and visible (UV-Vis) frequency range. In this frequency range, the planet reflects a fraction α of this energy where α is the albedo or reflectance of the planet in the UV-Vis range. In other words, the planet absorbs a fraction $1 - \alpha$ of the Sun's light, and reflects the rest. The power absorbed by the planet and its atmosphere is then:

$$P_{\text{abs}} = (1 - \alpha) P_{SE} \quad (3)$$

Even though the planet only absorbs as a circular area πR^2 , it emits equally in all directions as a sphere. If the planet were a perfect black body, it would emit according to the Stefan–Boltzmann law

$$P_{\text{emt bb}} = 4\pi R_E^2 \sigma T_E^4 \quad (4)$$

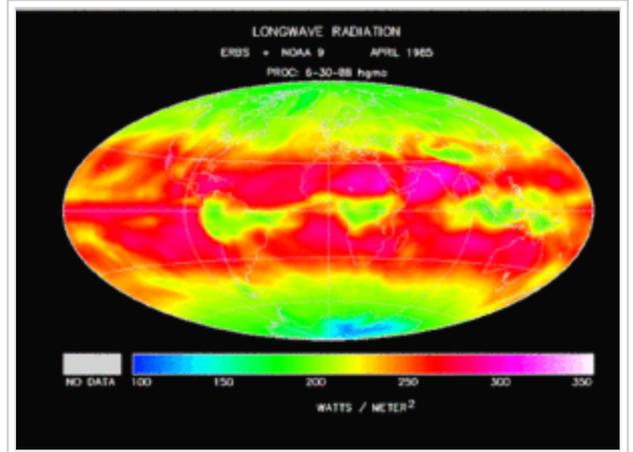
where T_E is the temperature of the planet. This temperature, calculated for the case of the planet acting as a black body by setting $P_{\text{abs}} = P_{\text{emt bb}}$, is known as the effective temperature. The actual temperature of the planet will likely be different, depending on its surface and atmospheric properties. Ignoring the atmosphere and greenhouse effect, the planet, since it is at a much lower temperature than the Sun, emits mostly in the infrared (IR) portion of the spectrum. In this frequency range, it emits $\bar{\epsilon}$ of the radiation that a black body would emit where $\bar{\epsilon}$ is the average emissivity in the IR range. The power emitted by the planet is then:

$$P_{\text{emt}} = \bar{\epsilon} P_{\text{emt bb}} \quad (5)$$

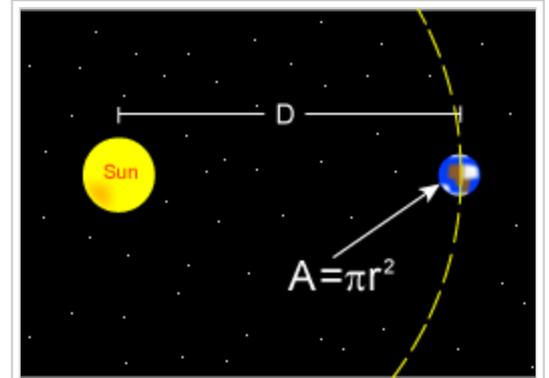
For a body in radiative exchange equilibrium with its surroundings, the rate at which it emits radiant energy is equal to the rate at which it absorbs it:^{[40][41]}

$$P_{\text{abs}} = P_{\text{emt}} \quad (6)$$

Substituting the expressions for solar and planet power in equations 1–6 and simplifying yields the estimated temperature of the planet, ignoring greenhouse effect, T_p :



Earth's longwave thermal radiation intensity, from clouds, atmosphere and ground



The Earth only has an absorbing area equal to a two dimensional disk, rather than the surface of a sphere.

$$T_P = T_S \sqrt{\frac{R_S \sqrt{\frac{1-\alpha}{\bar{\epsilon}}}}{2D}} \quad (7)$$

In other words, given the assumptions made, the temperature of a planet depends only on the surface temperature of the Sun, the radius of the Sun, the distance between the planet and the Sun, the albedo and the IR emissivity of the planet.

Notice that a gray (flat spectrum) ball where $(1 - \alpha) = \bar{\epsilon}$ comes to the same temperature as a black body no matter how dark or light gray .

Virtual temperature of Earth

Substituting the measured values for the Sun and Earth yields:

$$\begin{aligned} T_S &= 5778 \text{ K},^{[42]} \\ R_S &= 6.96 \times 10^8 \text{ m},^{[42]} \\ D &= 1.496 \times 10^{11} \text{ m},^{[42]} \\ \alpha &= 0.306^{[43]} \end{aligned}$$

With the average emissivity $\bar{\epsilon}$ set to unity, the effective temperature of the Earth is:

$$T_E = 254.356 \text{ K}$$

or $-18.8 \text{ }^\circ\text{C}$.

This is the temperature of the Earth if it radiated as a perfect black body in the infrared, assuming an unchanging albedo and ignoring greenhouse effects (which can raise the surface temperature of a body above what it would be if it were a perfect black body in all spectrums^[44]). The Earth in fact radiates not quite as a perfect black body in the infrared which will raise the estimated temperature a few degrees above the effective temperature. If we wish to estimate what the temperature of the Earth would be if it had no atmosphere, then we could take the albedo and emissivity of the Moon as a good estimate. The albedo and emissivity of the Moon are about 0.1054^[45] and 0.95^[46] respectively, yielding an estimated temperature of about 1.36 °C.

Estimates of the Earth's average albedo vary in the range 0.3–0.4, resulting in different estimated effective temperatures. Estimates are often based on the solar constant (total insolation power density) rather than the temperature, size, and distance of the Sun. For example, using 0.4 for albedo, and an insolation of 1400 W m⁻², one obtains an effective temperature of about 245 K.^[47] Similarly using albedo 0.3 and solar constant of 1372 W m⁻², one obtains an effective temperature of 255 K.^{[48][49][50]}

Cosmology

The cosmic microwave background radiation observed today is the most perfect black-body radiation ever observed in nature, with a temperature of about 2.7 K.^[51] It is a "snapshot" of the radiation at the time of decoupling between matter and radiation in the early universe. Prior to this time, most matter in the universe was in the form of an ionized plasma in thermal, though not full thermodynamic, equilibrium with radiation.

According to Kondepudi and Prigogine, at very high temperatures (above 10¹⁰ K; such temperatures existed in the very early universe), where the thermal motion separates protons and neutrons in spite of the strong nuclear forces, electron-positron pairs appear and disappear spontaneously and are in thermal equilibrium with electromagnetic radiation. These particles form a part of the black body spectrum, in addition to the electromagnetic radiation.^[52]

Doppler effect for a moving black body

The relativistic Doppler effect causes a shift in the frequency f of light originating from a source that is moving in relation to the observer, so that the wave is observed to have frequency f' :

$$f' = f \frac{1 - \frac{v}{c} \cos \theta}{\sqrt{1 - v^2/c^2}},$$

where v is the velocity of the source in the observer's rest frame, θ is the angle between the velocity vector and the observer-source direction measured in the reference frame of the source, and c is the speed of light.^[53] This can be simplified for the special cases of objects moving directly towards ($\theta = \pi$) or away ($\theta = 0$) from the observer, and for speeds much less than c .

Through Planck's law the temperature spectrum of a black body is proportionally related to the frequency of light and one may substitute the temperature (T) for the frequency in this equation.

For the case of a source moving directly towards or away from the observer, this reduces to

$$T' = T \sqrt{\frac{c - v}{c + v}}.$$

Here $v > 0$ indicates a receding source, and $v < 0$ indicates an approaching source.

This is an important effect in astronomy, where the velocities of stars and galaxies can reach significant fractions of c . An example is found in the cosmic microwave background radiation, which exhibits a dipole anisotropy from the Earth's motion relative to this black-body radiation field.

History

Balfour Stewart

In 1858, Balfour Stewart described his experiments on the thermal radiative emissive and absorptive powers of polished plates of various substances, compared with the powers of lamp-black surfaces, at the same temperature.^[23] Stewart chose lamp-black surfaces as his reference because of various previous experimental findings, especially those of Pierre Prevost and of John Leslie. He wrote "Lamp-black, which absorbs all the rays that fall upon it, and therefore possesses the greatest possible absorbing power, will possess also the greatest possible radiating power." More an experimenter than a logician, Stewart failed to point out that his statement presupposed an abstract general principle, that there exist either ideally in theory or really in nature bodies or surfaces that respectively have one and the same unique universal greatest possible absorbing power, likewise for radiating power, for every wavelength and equilibrium temperature.

Stewart measured radiated power with a thermo-pile and sensitive galvanometer read with a microscope. He was concerned with selective thermal radiation, which he investigated with plates of substances that radiated and absorbed selectively for different qualities of radiation rather than maximally for all qualities of radiation. He discussed the experiments in terms of rays which could be reflected and refracted, and which obeyed the Stokes-Helmholtz reciprocity principle (though he did not use an eponym for it). He did not in this paper mention that the qualities of the rays might be described by their wavelengths, nor did he use spectrally resolving apparatus such as prisms or diffraction gratings. His work was quantitative within these constraints. He made his measurements in a room temperature environment, and quickly so as to catch his bodies in a condition near the thermal equilibrium in which they had been prepared by heating to equilibrium with boiling water. His measurements confirmed that substances that emit and absorb selectively respect the principle of selective equality of emission and absorption at thermal equilibrium.

Stewart offered a theoretical proof that this should be the case separately for every selected quality of thermal radiation, but his mathematics was not rigorously valid.^[54] He made no mention of thermodynamics in this paper, though he did refer to conservation of *vis viva*. He proposed that his measurements implied that radiation was both absorbed and emitted by particles of matter throughout depths of the media in which it propagated. He applied the Helmholtz reciprocity principle to account for the material interface processes as distinct from the processes in the interior material. He did not postulate unrealizable perfectly black surfaces. He concluded that his experiments showed that in a cavity in thermal equilibrium, the heat radiated from any part of the interior bounding surface, no matter of what material it might be composed, was the same as would have been emitted from a surface of the same shape and position that would have been composed of lamp-black. He did not state explicitly that the lamp-black-coated bodies that he used as reference must have had a unique common spectral emittance function that depended on temperature in a unique way.

Gustav Kirchhoff

In 1859, not knowing of Stewart's work, Gustav Robert Kirchhoff reported the coincidence of the wavelengths of spectrally resolved lines of absorption and of emission of visible light. Importantly for thermal physics, he also observed that bright lines or dark lines were apparent depending on the temperature difference between emitter and absorber.^[55]

Kirchhoff then went on to consider bodies that emit and absorb heat radiation, in an opaque enclosure or cavity, in equilibrium at temperature T .

Here is used a notation different from Kirchhoff's. Here, the emitting power $E(T, i)$ denotes a dimensioned quantity, the total radiation emitted by a body labeled by index i at temperature T . The total absorption ratio $a(T, i)$ of that body is dimensionless, the ratio of absorbed to incident radiation in the cavity at temperature T . (In contrast with Balfour Stewart's, Kirchhoff's definition of his absorption ratio did not refer in particular to a lamp-black surface as the source of the incident radiation.) Thus the ratio $E(T, i) / a(T, i)$ of emitting power to absorption ratio is a dimensioned quantity, with the dimensions of emitting power, because $a(T, i)$ is dimensionless. Also here the wavelength-specific emitting power of the body at temperature T is denoted by $E(\lambda, T, i)$ and the wavelength-specific absorption ratio by $a(\lambda, T, i)$. Again, the ratio $E(\lambda, T, i) / a(\lambda, T, i)$ of emitting power to absorption ratio is a dimensioned quantity, with the dimensions of emitting power.

In a second report made in 1859, Kirchhoff announced a new general principle or law for which he offered a theoretical and mathematical proof, though he did not offer quantitative measurements of radiation powers.^[56] His theoretical proof was and still is considered by some writers to be invalid.^{[54][57]} His principle, however, has endured: it was that for heat rays of the same wavelength, in equilibrium at a given temperature, the wavelength-specific ratio of emitting power to absorption ratio has one and the same common value for all bodies that emit and absorb at that wavelength. In symbols, the law stated that the wavelength-specific ratio $E(\lambda, T, i) / a(\lambda, T, i)$ has one and the same value for all bodies, that is for all values of index i . In this report there was no mention of black bodies.

In 1860, still not knowing of Stewart's measurements for selected qualities of radiation, Kirchhoff pointed out that it was long established experimentally that for total heat radiation, of unselected quality, emitted and absorbed by a body in equilibrium, the dimensioned total radiation ratio $E(T, i) / a(T, i)$, has one and the same value common to all bodies, that is, for every value of the material index i .^[58] Again without measurements of radiative powers or other new experimental data, Kirchhoff then offered a fresh theoretical proof of his new principle of the universality of the value of the wavelength-specific ratio $E(\lambda, T, i) / a(\lambda, T, i)$ at thermal equilibrium. His fresh theoretical proof was and still is considered by some writers to be invalid.^{[54][57]}

But more importantly, it relied on a new theoretical postulate of "perfectly black bodies," which is the reason why one speaks of Kirchhoff's law. Such black bodies showed complete absorption in their infinitely thin most superficial surface. They correspond to Balfour Stewart's reference bodies, with internal radiation, coated with lamp-black. They were not the more realistic perfectly black bodies later considered by Planck. Planck's black

bodies radiated and absorbed only by the material in their interiors; their interfaces with contiguous media were only mathematical surfaces, capable neither of absorption nor emission, but only of reflecting and transmitting with refraction.^[59]

Kirchhoff's proof considered an arbitrary non-ideal body labeled i as well as various perfect black bodies labeled BB . It required that the bodies be kept in a cavity in thermal equilibrium at temperature T . His proof intended to show that the ratio $E(\lambda, T, i) / a(\lambda, T, i)$ was independent of the nature i of the non-ideal body, however partly transparent or partly reflective it was.

His proof first argued that for wavelength λ and at temperature T , at thermal equilibrium, all perfectly black bodies of the same size and shape have the one and the same common value of emissive power $E(\lambda, T, BB)$, with the dimensions of power. His proof noted that the dimensionless wavelength-specific absorption ratio $a(\lambda, T, BB)$ of a perfectly black body is by definition exactly 1. Then for a perfectly black body, the wavelength-specific ratio of emissive power to absorption ratio $E(\lambda, T, BB) / a(\lambda, T, BB)$ is again just $E(\lambda, T, BB)$, with the dimensions of power. Kirchhoff considered, successively, thermal equilibrium with the arbitrary non-ideal body, and with a perfectly black body of the same size and shape, in place in his cavity in equilibrium at temperature T . He argued that the flows of heat radiation must be the same in each case. Thus he argued that at thermal equilibrium the ratio $E(\lambda, T, i) / a(\lambda, T, i)$ was equal to $E(\lambda, T, BB)$, which may now be denoted $B_\lambda(\lambda, T)$, a continuous function, dependent only on λ at fixed temperature T , and an increasing function of T at fixed wavelength λ , at low temperatures vanishing for visible but not for longer wavelengths, with positive values for visible wavelengths at higher temperatures, which does not depend on the nature i of the arbitrary non-ideal body. (Geometrical factors, taken into detailed account by Kirchhoff, have been ignored in the foregoing.)

Thus Kirchhoff's law of thermal radiation can be stated: *For any material at all, radiating and absorbing in thermodynamic equilibrium at any given temperature T , for every wavelength λ , the ratio of emissive power to absorptive ratio has one universal value, which is characteristic of a perfect black body, and is an emissive power which we here represent by $B_\lambda(\lambda, T)$.* (For our notation $B_\lambda(\lambda, T)$, Kirchhoff's original notation was simply e .)^{[58][60][61][62][63][64]}

Kirchhoff announced that the determination of the function $B_\lambda(\lambda, T)$ was a problem of the highest importance, though he recognized that there would be experimental difficulties to be overcome. He supposed that like other functions that do not depend on the properties of individual bodies, it would be a simple function. Occasionally by historians that function $B_\lambda(\lambda, T)$ has been called "Kirchhoff's (emission, universal) function,"^{[65][66][67][68]} though its precise mathematical form would not be known for another forty years, till it was discovered by Planck in 1900. The theoretical proof for Kirchhoff's universality principle was worked on and debated by various physicists over the same time, and later.^[57] Kirchhoff stated later in 1860 that his theoretical proof was better than Balfour Stewart's, and in some respects it was so.^[54] Kirchhoff's 1860 paper did not mention the second law of thermodynamics, and of course did not mention the concept of entropy which had not at that time been established. In a more considered account in a book in 1862, Kirchhoff mentioned the connection of his law with Carnot's principle, which is a form of the second law.^[69]

According to Helge Kragh, "Quantum theory owes its origin to the study of thermal radiation, in particular to the "black-body" radiation that Robert Kirchhoff had first defined in 1859–1860."^[70]

See also

- Bolometer
- Color temperature
- Infrared thermometer
- Photon polarization
- Planck's law
- Pyrometry
- Rayleigh–Jeans law
- Thermography
- Sakuma–Hattori equation

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$$i_{w,t} = \frac{2hc^2}{w^5 (\exp(hc/wkt) - 1)}$$

$$i[w, t] = 2 * h * c^2 / (w^5 * (\text{Exp}[h * c / (w * k * t)] - 1))$$

The number of photons/sec/area is: $N = \int \frac{2 * \pi * i[w, t]}{(h * c / w)} \cdot \{w, 390 * 10^{-9}, 750 * 10^{-9}\} = 0.0244173...$

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External links

- Calculating Black-body Radiation Interactive calculator with Doppler Effect. Includes most systems of units.
- Color-to-Temperature demonstration at Academo.org
- Cooling Mechanisms for Human Body – From Hyperphysics
- Descriptions of radiation emitted by many different objects
- Black-Body Emission Applet
- "Blackbody Spectrum" by Jeff Bryant, Wolfram Demonstrations Project, 2007.

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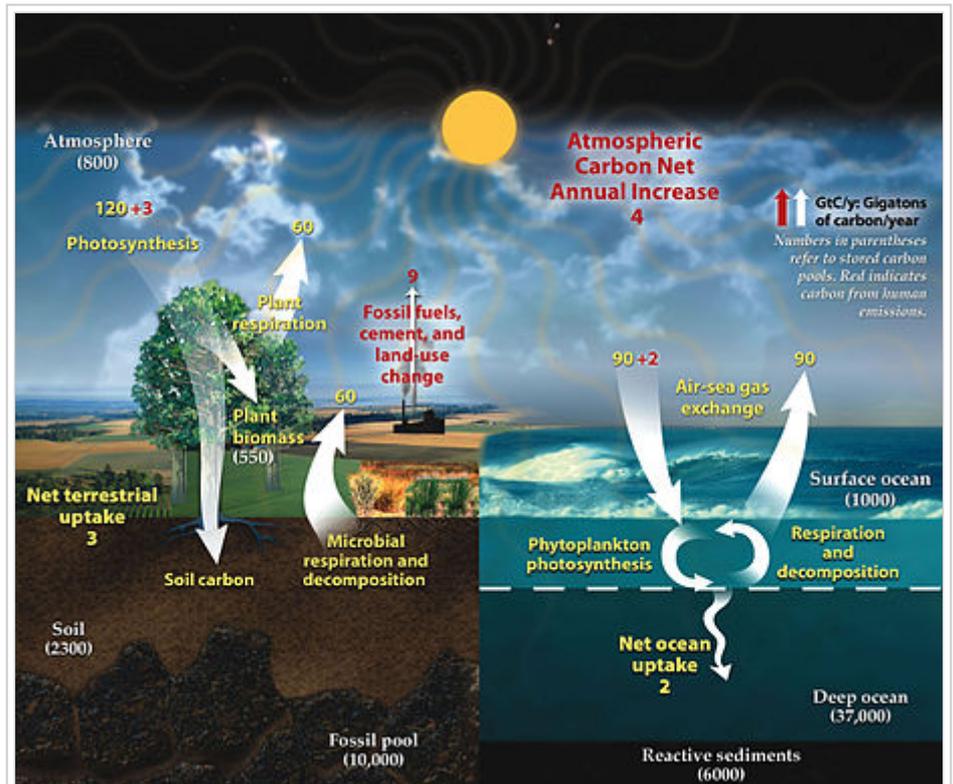
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Carbon cycle

From Wikipedia, the free encyclopedia

The **carbon cycle** is the biogeochemical cycle by which carbon is exchanged among the biosphere, pedosphere, geosphere, hydrosphere, and atmosphere of the Earth. Carbon is the main component of biological compounds as well as a major component of many minerals such as limestone. Along with the nitrogen cycle and the water cycle, the carbon cycle comprises a sequence of events that are key to make Earth capable of sustaining life. It describes the movement of carbon as it is recycled and reused throughout the biosphere, as well as long-term processes of carbon sequestration to and release from carbon sinks.

The *global carbon budget* is the balance of the exchanges (incomes and losses) of carbon between the carbon reservoirs or between one specific loop (e.g., atmosphere and biosphere) of the carbon cycle. An examination of the carbon budget of a pool or reservoir can provide information about whether the pool or reservoir is functioning as a source or sink for carbon dioxide. The carbon cycle was initially discovered by Joseph Priestley and Antoine Lavoisier, and popularized by Humphry Davy.^[1]



This diagram of the fast carbon cycle shows the movement of carbon between land, atmosphere, and oceans in billions of tons per year. Yellow numbers are natural fluxes, red are human contributions, white indicate stored carbon. Note this diagram does not account for volcanic and tectonic activity which also sequesters and releases carbon.

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 - 2.3 Oceans
 - 2.4 Geological carbon cycle
 - 2.5 Human influence
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Global climate

Carbon exists in various forms in the atmosphere. Carbon dioxide (CO₂) and methane (CH₄) are partly responsible for the greenhouse effect and among the most important human-contributed greenhouse gases.^[2]

In the past two centuries, human activities have altered the global carbon cycle, most significantly in the atmosphere. Although carbon dioxide levels have changed naturally over the past several thousand years, human emissions of carbon dioxide into the atmosphere have created unnatural fluctuations.^[2] Changes in the amount of atmospheric CO₂ are considerably altering weather patterns and indirectly influencing oceanic chemistry. Current carbon dioxide levels in the atmosphere exceed measurements from the last 1,000 years and levels are rising quickly,^[3] making it of critical importance to better understand how the carbon cycle works and what its effects are on the global climate.^[2]

Main components

The global carbon cycle is now usually divided into the following major reservoirs of carbon interconnected by pathways of exchange:

- The atmosphere
- The terrestrial biosphere
- The oceans, including dissolved inorganic carbon and living and non-living marine biota
- The sediments, including fossil fuels, fresh water systems and non-living organic material.
- The Earth's interior, carbon from the Earth's mantle and crust. These carbon stores interact with the other components through geological processes

The carbon exchanges between reservoirs occur as the result of various chemical, physical, geological, and biological processes. The ocean contains the largest active pool of carbon near the surface of the Earth.^[2] The natural flows of carbon between the atmosphere, ocean, terrestrial ecosystems, and sediments is fairly balanced, so that carbon levels would be roughly stable without human influence.^{[4][5]}

Atmosphere

Carbon in the Earth's atmosphere exists in two main forms: carbon dioxide and methane. Both of these gases absorb and retain heat in the atmosphere and are partially responsible for the greenhouse effect. Methane produces a larger greenhouse effect per volume as compared to carbon dioxide, but it exists in much lower concentrations and is more short-lived than carbon dioxide, making carbon dioxide the more important greenhouse gas of the two.^[6]

Carbon dioxide is removed from the atmosphere primarily through photosynthesis and enters the terrestrial and oceanic biospheres. Carbon dioxide also dissolves directly from the atmosphere into bodies of water (oceans, lakes, etc.), as well as dissolving in precipitation as raindrops fall through the atmosphere. When dissolved in water, carbon dioxide reacts with water molecules and forms carbonic acid, which contributes to ocean acidity. It can then be absorbed by rocks through weathering. It also can acidify other surfaces it touches or be washed into the ocean.^[7]

Carbon pools in the major reservoirs on earth.^[2]

Pool	Quantity (gigatons)
Atmosphere	720
Oceans (total)	38,400
Total inorganic	37,400
Total organic	1,000
Surface layer	670
Deep layer	36,730
Lithosphere	
Sedimentary carbonates	> 60,000,000
Kerogens	15,000,000
Terrestrial biosphere (total)	2,000
Living biomass	600 - 1,000
Dead biomass	1,200
Aquatic biosphere	1 - 2
Fossil fuels (total)	4,130
Coal	3,510
Oil	230
Gas	140
Other (peat)	250

Human activities over the past two centuries have significantly increased the amount of carbon in the atmosphere, mainly in the form of carbon dioxide, both by modifying ecosystems' ability to extract carbon dioxide from the atmosphere and by emitting it directly, e.g., by burning fossil fuels and manufacturing concrete.^[2]

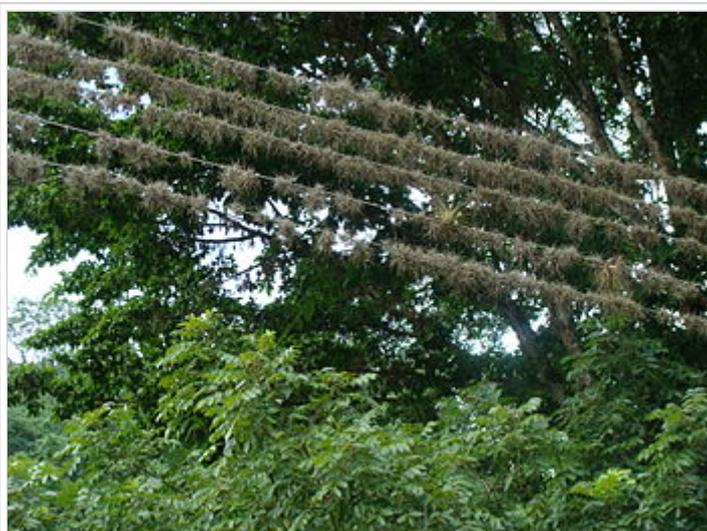
Terrestrial biosphere

The terrestrial biosphere includes the organic carbon in all land-living organisms, both alive and dead, as well as carbon stored in soils. About 500 gigatons of carbon are stored above ground in plants and other living organisms,^[4] while soil holds approximately 1,500 gigatons of carbon.^[8] Most carbon in the terrestrial biosphere is organic carbon,^[9] while about a third of soil carbon is stored in inorganic forms, such as calcium carbonate.^[10] Organic carbon is a major component of all organisms living on earth. Autotrophs extract it from the air in the form of carbon dioxide, converting it into organic carbon, while heterotrophs receive carbon by consuming other organisms.

Because carbon uptake in the terrestrial biosphere is dependent on biotic factors, it follows a diurnal and seasonal cycle. In CO₂ measurements, this feature is apparent in the Keeling curve. It is strongest in the northern hemisphere, because this hemisphere has more land mass than the southern hemisphere and thus more room for ecosystems to absorb and emit carbon.

Carbon leaves the terrestrial biosphere in several ways and on different time scales. The combustion or respiration of organic carbon releases it rapidly into the atmosphere. It can also be exported into the oceans through rivers or remain sequestered in soils in the form of inert carbon. Carbon stored in soil can remain there for up to thousands of years before being washed into rivers by erosion or released into the atmosphere through soil respiration. Between 1989 and 2008 soil respiration increased by about 0.1% per year.^[11] In 2008, the global total of CO₂ released from the soil reached roughly 98 billion tonnes, about 10 times more carbon than humans are now putting into the atmosphere each year by burning fossil fuel. There are a few plausible explanations for this trend, but the most likely explanation is that increasing temperatures have increased rates of decomposition of soil organic matter, which has increased the flow of CO₂. The length of carbon sequestering in soil is dependent on local climatic conditions and thus changes in the course of climate change. From pre-industrial era to 2010, the terrestrial biosphere represented a net source of atmospheric CO₂ prior to 1940, switching subsequently to a net sink.^[12]

Oceans



Epiphytes on electric wires. This kind of plant takes both CO₂ and water from the atmosphere for living and growing.



A portable soil respiration system measuring soil CO₂ flux

Oceans contain the greatest quantity of actively cycled carbon in this world and are second only to the lithosphere in the amount of carbon they store.^[2] The oceans' surface layer holds large amounts of dissolved inorganic carbon that is exchanged rapidly with the atmosphere. The deep layer's concentration of dissolved inorganic carbon (DIC) is about 15% higher than that of the surface layer.^[13] DIC is stored in the deep layer for much longer periods of time.^[4] Thermohaline circulation exchanges carbon between these two layers.^[2]

Carbon enters the ocean mainly through the dissolution of atmospheric carbon dioxide, which is converted into carbonate. It can also enter the oceans through rivers as dissolved organic carbon. It is converted by organisms into organic carbon through photosynthesis and can either be exchanged throughout the food chain or precipitated into the ocean's deeper, more carbon rich layers as dead soft tissue or in shells as calcium carbonate. It circulates in this layer for long periods of time before either being deposited as sediment or, eventually, returned to the surface waters through thermohaline circulation.^[4]

Oceanic absorption of CO₂ is one of the most important forms of carbon sequestering limiting the human-caused rise of carbon dioxide in the atmosphere. However, this process is limited by a number of factors. Because the rate of CO₂ dissolution in the ocean is dependent on the weathering of rocks and this process takes place slower than current rates of human greenhouse gas emissions, ocean CO₂ uptake will decrease in the future.^[2] CO₂ absorption also makes water more acidic, which affects ocean biosystems. The projected rate of increasing oceanic acidity could slow the biological precipitation of calcium carbonates, thus decreasing the ocean's capacity to absorb carbon dioxide.^{[14][15]}

Geological carbon cycle

The geologic component of the carbon cycle operates slowly in comparison to the other parts of the global carbon cycle. It is one of the most important determinants of the amount of carbon in the atmosphere, and thus of global temperatures.^[16]

Most of the earth's carbon is stored inertly in the earth's lithosphere.^[2] Much of the carbon stored in the earth's mantle was stored there when the earth formed.^[17] Some of it was deposited in the form of organic carbon from the biosphere.^[18] Of the carbon stored in the geosphere, about 80% is limestone and its derivatives, which form from the sedimentation of calcium carbonate stored in the shells of marine organisms. The remaining 20% is stored as kerogens formed through the sedimentation and burial of terrestrial organisms under high heat and pressure. Organic carbon stored in the geosphere can remain there for millions of years.^[16]

Carbon can leave the geosphere in several ways. Carbon dioxide is released during the metamorphosis of carbonate rocks when they are subducted into the earth's mantle. This carbon dioxide can be released into the atmosphere and ocean through volcanoes and hotspots.^[17] It can also be removed by humans through the direct extraction of kerogens in the form of fossil fuels. After extraction, fossil fuels are burned to release energy, thus emitting the carbon they store into the atmosphere.

Human influence

Since the industrial revolution, human activity has modified the carbon cycle by changing its components' functions and directly adding carbon to the atmosphere.^[2]

The largest human impact on the carbon cycle is through direct emissions from burning fossil fuels, which transfers carbon from the geosphere into the atmosphere. The rest of this increase is caused mostly by changes in land-use, particularly deforestation.

Another direct human impact on the carbon cycle is the chemical process of calcination of limestone for clinker production, which releases CO₂.^[23] Clinker is an industrial precursor of cement.

Humans also influence the carbon cycle indirectly by changing the terrestrial and oceanic biosphere. Over the past several centuries, direct and indirect human-caused land use and land cover change (LUCC) has led to the loss of biodiversity, which lowers ecosystems' resilience to environmental stresses and decreases their ability to remove carbon from the atmosphere. More directly, it often leads to the release of carbon from terrestrial ecosystems into the atmosphere. Deforestation for agricultural purposes removes forests, which hold large amounts of carbon, and replaces them, generally with agricultural or urban areas. Both of these replacement land cover types store comparatively small amounts of carbon, so that the net product of the process is that more carbon stays in the atmosphere.

Other human-caused changes to the environment change ecosystems' productivity and their ability to remove carbon from the atmosphere. Air pollution, for example, damages plants and soils, while many agricultural and land use practices lead to higher erosion rates, washing carbon out of soils and decreasing plant productivity.

Humans also affect the oceanic carbon cycle. Current trends in climate change lead to higher ocean temperatures, thus modifying ecosystems. Also, acid rain and polluted runoff from agriculture and industry change the ocean's chemical composition. Such changes can have dramatic effects on highly sensitive ecosystems such as coral reefs, thus limiting the ocean's ability to absorb carbon from the atmosphere on a regional scale and reducing oceanic biodiversity globally.

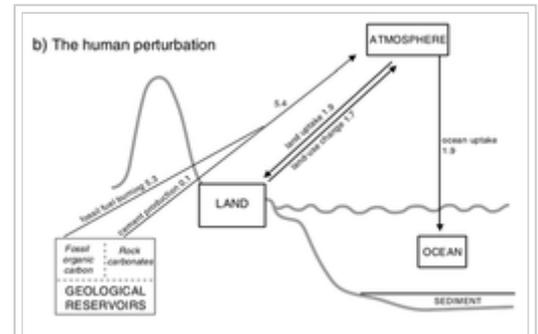
Arctic methane emissions indirectly caused by anthropogenic global warming also affect the carbon cycle, and contribute to further warming in what is known as climate change feedback.

On 12 November 2015, NASA scientists reported that human-made carbon dioxide (CO₂) continues to increase above levels not seen in hundreds of thousands of years: currently, about half of the carbon dioxide released from the burning of fossil fuels remains in the atmosphere and is not absorbed by vegetation and the oceans.^{[19][20][21][22]}

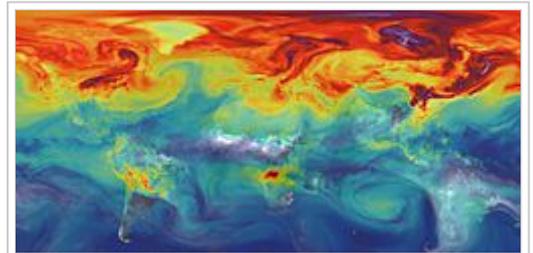
See also

- Biochar
- Calvin cycle
- Carbon cycle re-balancing
- Carbon dioxide in Earth's atmosphere
- Carbon footprint
- Deficit irrigation
- Nitrogen cycle
- Ocean acidification
- Permafrost carbon cycle
- Snowball Earth and the "Slow carbon cycle"
- Soil plant atmosphere continuum
- Soil regeneration

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Human activity since the industrial era has changed the balance in the natural carbon cycle. Units are in gigatons.^[4]



CO₂ in Earth's atmosphere if *half* of global-warming emissions are *not* absorbed.^{[19][20][21][22]} (NASA computer simulation).

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External links

- Carbon Cycle Science Program – an interagency partnership.
- NOAA's Carbon Cycle Greenhouse Gases Group
- Global Carbon Project – initiative of the Earth System Science Partnership
- UNEP – The present carbon cycle – Climate Change carbon levels and flows
- NASA's Orbiting Carbon Observatory
- CarboSchools, a European website with many resources to study carbon cycle in secondary schools.
- Carbon and Climate, an educational website with a carbon cycle applet for modeling your own projection.



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Climate model

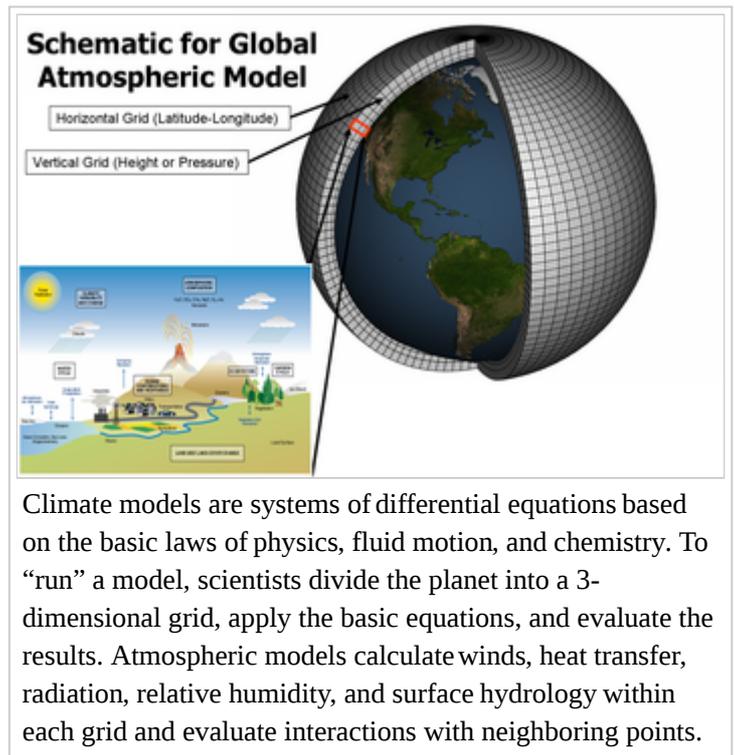
From Wikipedia, the free encyclopedia

Climate models use quantitative methods to simulate the interactions of the important drivers of climate, including atmosphere, oceans, land surface and ice. They are used for a variety of purposes from study of the dynamics of the climate system to projections of future climate.

All climate models take account of incoming energy from the sun as short wave electromagnetic radiation, chiefly visible and short-wave (near) infrared, as well as outgoing long wave (far) infrared electromagnetic. Any imbalance results in a change in temperature.

Models vary in complexity:

- A simple radiant heat transfer model treats the earth as a single point and averages outgoing energy
- This can be expanded vertically (radiative-convective models) and/or horizontally
- Finally, (coupled) atmosphere–ocean–sea ice **global climate models** solve the full equations for mass and energy transfer and radiant exchange.
- Box models can treat flows across and within ocean basins.
- Other types of modelling can be interlinked, such as land use, allowing researchers to predict the interaction between climate and ecosystems.



Climate models are systems of differential equations based on the basic laws of physics, fluid motion, and chemistry. To “run” a model, scientists divide the planet into a 3-dimensional grid, apply the basic equations, and evaluate the results. Atmospheric models calculate winds, heat transfer, radiation, relative humidity, and surface hydrology within each grid and evaluate interactions with neighboring points.

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Box models

Box models are simplified versions of complex systems, reducing them to boxes (or reservoirs) linked by fluxes. The boxes are assumed to be mixed homogeneously. Within a given box, the concentration of any chemical species is therefore uniform. However, the abundance of a species within a given box may vary as a function of time due to the input to (or loss from) the box or due to the production, consumption or decay of this species within the box.

Simple box models, i.e. box model with a small number of boxes whose properties (e.g. their volume) do not change with time, are often useful to derive analytical formulas describing the dynamics and steady-state abundance of a species. More complex box models are usually solved using numerical techniques.

Box models are used extensively to model environmental systems or ecosystems and in studies of ocean circulation and the carbon cycle.^[1]

Zero-dimensional models

A very simple model of the radiative equilibrium of the Earth is

$$(1 - a)S\pi r^2 = 4\pi r^2 \epsilon \sigma T^4$$

where

- the left hand side represents the incoming energy from the Sun
- the right hand side represents the outgoing energy from the Earth, calculated from the Stefan-Boltzmann law assuming a model-fictive temperature, T , sometimes called the 'equilibrium temperature of the Earth', that is to be found,

and

- S is the solar constant – the incoming solar radiation per unit area—about $1367 \text{ W}\cdot\text{m}^{-2}$
- a is the Earth's average albedo, measured to be 0.3.^{[2][3]}
- r is Earth's radius—approximately $6.371 \times 10^6 \text{ m}$
- π is the mathematical constant (3.141...)
- σ is the Stefan-Boltzmann constant—approximately $5.67 \times 10^{-8} \text{ J}\cdot\text{K}^{-4}\cdot\text{m}^{-2}\cdot\text{s}^{-1}$
- ϵ is the effective emissivity of earth, about 0.612

The constant πr^2 can be factored out, giving

$$(1 - a)S = 4\epsilon \sigma T^4$$

Solving for the temperature,

$$T = \sqrt[4]{\frac{(1 - a)S}{4\epsilon \sigma}}$$

This yields an apparent effective average earth temperature of 288 K (15 °C; 59 °F).^[4] This is because the above equation represents the effective *radiative* temperature of the Earth (including the clouds and atmosphere). The use of effective emissivity and albedo account for the greenhouse effect.

This very simple model is quite instructive, and the only model that could fit on a page. For example, it easily determines the effect on average earth temperature of changes in solar constant or change of albedo or effective earth emissivity.

The average emissivity of the earth is readily estimated from available data. The emissivities of terrestrial surfaces are all in the range of 0.96 to 0.99^{[5][6]} (except for some small desert areas which may be as low as 0.7). Clouds, however, which cover about half of the earth's surface, have an average emissivity of about 0.5^[7] (which must be reduced by the fourth power of the ratio of cloud absolute temperature to average earth absolute temperature) and an average cloud temperature of about 258 K (−15 °C; 5 °F).^[8] Taking all this properly into account results in an effective earth emissivity of about 0.64 (earth average temperature 285 K (12 °C; 53 °F)).

This simple model readily determines the effect of changes in solar output or change of earth albedo or effective earth emissivity on average earth temperature. It says nothing, however about what might cause these things to change. Zero-dimensional models do not address the temperature distribution on the earth or the factors that move energy about the earth.

Radiative-convective models

The zero-dimensional model above, using the solar constant and given average earth temperature, determines the effective earth emissivity of long wave radiation emitted to space. This can be refined in the vertical to a one-dimensional radiative-convective model, which considers two processes of energy transport:

- upwelling and downwelling radiative transfer through atmospheric layers that both absorb and emit infrared radiation
- upward transport of heat by convection (especially important in the lower troposphere).

The radiative-convective models have advantages over the simple model: they can determine the effects of varying greenhouse gas concentrations on effective emissivity and therefore the surface temperature. But added parameters are needed to determine local emissivity and albedo and address the factors that move energy about the earth.

Effect of ice-albedo feedback on global sensitivity in a one-dimensional radiative-convective climate model.^{[9][10][11]}

Higher-dimension models

The zero-dimensional model may be expanded to consider the energy transported horizontally in the atmosphere. This kind of model may well be zonally averaged. This model has the advantage of allowing a rational dependence of local albedo and emissivity on temperature – the poles can be allowed to be icy and the equator warm – but the lack of true dynamics means that horizontal transports have to be specified.^[12]

EMICs (Earth-system models of intermediate complexity)

Depending on the nature of questions asked and the pertinent time scales, there are, on the one extreme, conceptual, more inductive models, and, on the other extreme, general circulation models operating at the highest spatial and temporal resolution currently feasible. Models of intermediate complexity bridge the gap. One example is the Climber-3 model. Its atmosphere is a 2.5-dimensional statistical-dynamical model with $7.5^\circ \times 22.5^\circ$ resolution and time step of half a day; the ocean is MOM-3 (Modular Ocean Model) with a $3.75^\circ \times 3.75^\circ$ grid and 24 vertical levels.^[13]

GCMs (global climate models or general circulation models)

General Circulation Models (GCMs) discretise the equations for fluid motion and energy transfer and integrate these over time. Unlike simpler models, GCMs divide the atmosphere and/or oceans into grids of discrete "cells", which represent computational units. Unlike simpler models which make mixing assumptions, processes internal to a cell—such as convection—that occur on scales too small to be resolved directly are parameterised at the cell level, while other functions govern the interface between cells.

Atmospheric GCMs (AGCMs) model the atmosphere and impose sea surface temperatures as boundary conditions. Coupled atmosphere-ocean GCMs (AOGCMs, e.g. HadCM3, EdGCM, GFDL CM2.X, ARPEGE-Climat)^[14] combine the two models. The first general circulation climate model that combined both oceanic and atmospheric processes was developed in the late 1960s at the NOAA Geophysical Fluid Dynamics Laboratory^[15] AOGCMs represent the pinnacle of complexity in climate models and internalise as many

processes as possible. However, they are still under development and uncertainties remain. They may be coupled to models of other processes, such as the carbon cycle, so as to better model feedback effects. Such integrated multi-system models are sometimes referred to as either "earth system models" or "global climate models."

Research and development

There are three major types of institution where climate models are developed, implemented and used:

- National meteorological services. Most national weather services have a climatology section.
- Universities. Relevant departments include atmospheric sciences, meteorology, climatology, and geography.
- National and international research laboratories. Examples include the National Center for Atmospheric Research (NCAR, in Boulder, Colorado, USA), the Geophysical Fluid Dynamics Laboratory (GFDL, in Princeton, New Jersey, USA), the Hadley Centre for Climate Prediction and Research (in Exeter, UK), the Max Planck Institute for Meteorology in Hamburg, Germany, or the Laboratoire des Sciences du Climat et de l'Environnement (LSCE), France, to name but a few.

The World Climate Research Programme (WCRP), hosted by the World Meteorological Organization (WMO), coordinates research activities on climate modelling worldwide.

A 2012 U.S. National Research Council report discussed how the large and diverse U.S. climate modeling enterprise could evolve to become more unified.^[16] Efficiencies could be gained by developing a common software infrastructure shared by all U.S. climate researchers, and holding an annual climate modeling forum, the report found.^[17]

See also

- General circulation model
- Atmospheric Radiation Measurement (ARM) (in the US)
- Climateprediction.net
- GFDL CM2.X
- Numerical Weather Prediction
- Tropical cyclone prediction model

Climate models on the web

- Dapper/DChart — plot and download model data referenced by the Fourth Assessment Report (AR4) of the Intergovernmental Panel on Climate Change. (No longer available)
- NCAR/UCAR Community Climate System Model (CCSM)
- Do it yourself climate prediction
- Primary research GCM developed by NASA/GISS (Goddard Institute for Space Studies)
- Original NASA/GISS global climate model (GCM) with a user-friendly interface for PCs and Macs
- CCCma model info and interface to retrieve model data
- NOAA/Geophysical Fluid Dynamics Laboratory CM2 global climate model info and model output data files
- University of Victoria Global climate model, free for download. Leading researcher was a contributing author to an IPCC report on climate change.
- vimeo.com/user12523377/videos Visualizations of climate models of ETH Zurich

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External links

- (IPCC 2001 section 8.3) — on model hierarchy
- (IPCC 2001 section 8) — information on coupled GCM's
- Coupled Model Intercomparison Project
- On the Radiative and Dynamical Feedbacks over the Equatorial Pacific Cold Tongue
- Basic Radiation Calculations — The Discovery of Global Warming
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The Wikibook *Historical Geology* has a page on the topic of: **Climate models**

- Climate Modeling 101 website by the U.S. National Research Council — This site is a primer on how climate models work. The information is based on expert, consensus reports from the U.S. National Research Council's Board on Atmospheric Sciences and Climate. The most recent is *A National Strategy for Advancing Climate Modeling*.

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Categories: [Climate change science](#) | [Numerical climate and weather models](#) | [Climate modeling](#)

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Cryosphere

From Wikipedia, the free encyclopedia

The **cryosphere** (from the Greek κρύος *kryos*, "cold", "frost" or "ice" and σφαῖρα *sphaira*, "globe, ball"^[1]) is those portions of Earth's surface where water is in solid form, including sea ice, lake ice, river ice, snow cover, glaciers, ice caps, ice sheets, and frozen ground (which includes permafrost). Thus, there is a wide overlap with the hydrosphere. The cryosphere is an integral part of the global climate system with important linkages and feedbacks generated through its influence on surface energy and moisture fluxes, clouds, precipitation, hydrology, atmospheric and oceanic circulation. Through these feedback processes, the cryosphere plays a significant role in the global climate and in climate model response to global changes. The term deglaciation describes the retreat of cryospheric features. Cryology is the study of cryospheres.



Overview of the Cryosphere and its larger components, from the UN Environment Programme Global Outlook for Ice and Snow.

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Structure

Frozen water is found on the Earth's surface primarily as snow cover, freshwater ice in lakes and rivers, sea ice, glaciers, ice sheets, and frozen ground and permafrost (permanently frozen ground). The residence time of water in each of these cryospheric sub-systems varies widely. Snow cover and freshwater ice are essentially seasonal, and most sea ice, except for ice in the central Arctic, lasts only a few years if it is not seasonal. A given water particle in glaciers, ice sheets, or ground ice, however, may remain frozen for 10-100,000 years or longer, and deep ice in parts of East Antarctica may have an age approaching 1 million years.

Most of the world's ice volume is in Antarctica, principally in the East Antarctic Ice Sheet. In terms of areal extent, however, Northern Hemisphere winter snow and ice extent comprise the largest area, amounting to an average 23% of hemispheric surface area in January. The large areal extent and the important climatic roles of snow and ice, related to their unique physical properties, indicate that the ability to observe and model snow and ice-cover extent, thickness, and physical properties (radiative and thermal properties) is of particular significance for climate research.

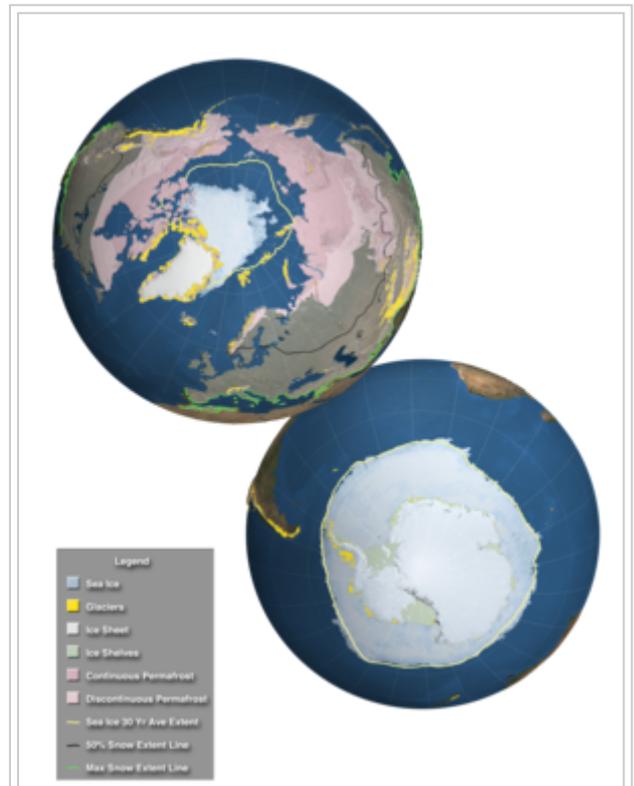
There are several fundamental physical properties of snow and ice that modulate energy exchanges between the surface and the atmosphere. The most important properties are the surface reflectance (albedo), the ability to transfer heat (thermal diffusivity), and the ability to change state (latent heat). These physical properties, together with surface roughness, emissivity, and dielectric characteristics, have important implications for

observing snow and ice from space. For example, surface roughness is often the dominant factor determining the strength of radar backscatter.^[2] Physical properties such as crystal structure, density, length, and liquid water content are important factors affecting the transfers of heat and water and the scattering of microwave energy.

The surface reflectance of incoming solar radiation is important for the surface energy balance (SEB). It is the ratio of reflected to incident solar radiation, commonly referred to as albedo. Climatologists are primarily interested in albedo integrated over the shortwave portion of the electromagnetic spectrum (~300 to 3500 nm), which coincides with the main solar energy input. Typically, albedo values for non-melting snow-covered surfaces are high (~80-90%) except in the case of forests. The higher albedos for snow and ice cause rapid shifts in surface reflectivity in autumn and spring in high latitudes, but the overall climatic significance of this increase is spatially and temporally modulated by cloud cover. (Planetary albedo is determined principally by cloud cover, and by the small amount of total solar radiation received in high latitudes during winter months.) Summer and autumn are times of high-average cloudiness over the Arctic Ocean so the albedo feedback associated with the large seasonal changes in sea-ice extent is greatly reduced. Groisman *et al.* (1994a) observed that snow cover exhibited the greatest influence on the Earth radiative balance in the spring (April to May) period when incoming solar radiation was greatest over snow-covered areas.^[3]

The thermal properties of cryospheric elements also have important climatic consequences. Snow and ice have much lower thermal diffusivities than air. Thermal diffusivity is a measure of the speed at which temperature waves can penetrate a substance. Snow and ice are many orders of magnitude less efficient at diffusing heat than air. Snow cover insulates the ground surface, and sea ice insulates the underlying ocean, decoupling the surface-atmosphere interface with respect to both heat and moisture fluxes. The flux of moisture from a water surface is eliminated by even a thin skin of ice, whereas the flux of heat through thin ice continues to be substantial until it attains a thickness in excess of 30 to 40 cm. However, even a small amount of snow on top of the ice will dramatically reduce the heat flux and slow down the rate of ice growth. The insulating effect of snow also has major implications for the hydrological cycle. In non-permafrost regions, the insulating effect of snow is such that only near-surface ground freezes and deep-water drainage is uninterrupted.^[4]

While snow and ice act to insulate the surface from large energy losses in winter, they also act to retard warming in the spring and summer because of the large amount of energy required to melt ice (the latent heat of fusion, 3.34×10^5 J/kg at 0 °C). However, the strong static stability of the atmosphere over areas of extensive snow or ice tends to confine the immediate cooling effect to a relatively shallow layer, so that associated atmospheric anomalies are usually short-lived and local to regional in scale.^[5] In some areas of the world such



This high resolution image, designed for the Fifth Assessment Report of the IPCC, shows the extent of the regions affected by components of the cryosphere around the world. Over land, continuous permafrost is shown in a dark pink while discontinuous permafrost is shown in a lighter shade of pink. Over much of the northern hemisphere's land area, a semi-transparent white veil depicts the regions that were affected by snowfall at least one day during the period 2000-2012. The bright green line along the southern border of this region shows the maximum snow extent while a black line across the North America, Europe and Asia shows the 50% snow extent line. Glaciers are shown as small golden dots in mountainous areas and in the far northern and southern latitudes. Over the water, ice shelves are shown around Antarctica along with sea ice surrounding the ice shelves. Sea ice is also shown at the North Pole. For both poles, the 30 year average sea ice extent is shown by a yellow outline. In addition, the ice sheets of Greenland and Antarctica are clearly visible.

as Eurasia, however, the cooling associated with a heavy snowpack and moist spring soils is known to play a role in modulating the summer monsoon circulation.^[6] Gutzler and Preston (1997) recently presented evidence for a similar snow-summer circulation feedback over the southwestern United States.^[7]

The role of snow cover in modulating the monsoon is just one example of a short-term cryosphere-climate feedback involving the land surface and the atmosphere. From Figure 1 it can be seen that there are numerous cryosphere-climate feedbacks in the global climate system. These operate over a wide range of spatial and temporal scales from local seasonal cooling of air temperatures to hemispheric-scale variations in ice sheets over time-scales of thousands of years. The feedback mechanisms involved are often complex and incompletely understood. For example, Curry *et al.* (1995) showed that the so-called “simple” sea ice-albedo feedback involved complex interactions with lead fraction, melt ponds, ice thickness, snow cover, and sea-ice extent.

Snow

Snow cover has the second-largest areal extent of any component of the cryosphere, with a mean maximum areal extent of approximately 47 million km². Most of the Earth’s snow-covered area (SCA) is located in the Northern Hemisphere, and temporal variability is dominated by the seasonal cycle; Northern Hemisphere snow-cover extent ranges from 46.5 million km² in January to 3.8 million km² in August.^[8] North American winter SCA has exhibited an increasing trend over much of this century (Brown and Goodison 1996; Hughes *et al.* 1996) largely in response to an increase in precipitation.^[9] However, the available satellite data show that the hemispheric winter snow cover has exhibited little interannual variability over the 1972-1996 period, with a coefficient of variation (COV=s.d./mean) for January Northern Hemisphere snow cover of < 0.04. According to Groisman *et al.* (1994a) Northern Hemisphere spring snow cover should exhibit a decreasing trend to explain an observed increase in Northern Hemisphere spring air temperatures this century. Preliminary estimates of SCA from historical and reconstructed in situ snow-cover data suggest this is the case for Eurasia, but not for North America, where spring snow cover has remained close to current levels over most of this century.^[10] Because of the close relationship observed between hemispheric air temperature and snow-cover extent over the period of satellite data (IPCC 1996), there is considerable interest in monitoring Northern Hemisphere snow-cover extent for detecting and monitoring climate change.

Snow cover is an extremely important storage component in the water balance, especially seasonal snowpacks in mountainous areas of the world. Though limited in extent, seasonal snowpacks in the Earth’s mountain ranges account for the major source of the runoff for stream flow and groundwater recharge over wide areas of the midlatitudes. For example, over 85% of the annual runoff from the Colorado River basin originates as snowmelt. Snowmelt runoff from the Earth’s mountains fills the rivers and recharges the aquifers that over a billion people depend on for their water resources. Further, over 40% of the world’s protected areas are in mountains, attesting to their value both as unique ecosystems needing protection and as recreation areas for humans. Climate warming is expected to result in major changes to the partitioning of snow and rainfall, and to the timing of snowmelt, which will have important implications for water use and management. These changes also involve potentially important decadal and longer time-scale feedbacks to the climate system through temporal and spatial changes in soil moisture and runoff to the oceans.(Walsh 1995). Freshwater fluxes from the snow cover into the marine environment may be important, as the total flux is probably of the same magnitude as desalinated ridging and rubble areas of sea ice.^[11] In addition, there is an associated pulse of precipitated pollutants which accumulate over the Arctic winter in snowfall and are released into the ocean upon ablation of the sea-ice .

Sea ice

Sea ice covers much of the polar oceans and forms by freezing of sea water. Satellite data since the early 1970s reveal considerable seasonal, regional, and interannual variability in the sea-ice covers of both hemispheres. Seasonally, sea-ice extent in the Southern Hemisphere varies by a factor of 5, from a minimum of 3-4 million km² in February to a maximum of 17-20 million km² in September.^{[12][13]} The seasonal variation is much less in the Northern Hemisphere where the confined nature and high latitudes of the Arctic Ocean result in a much

larger perennial ice cover, and the surrounding land limits the equatorward extent of wintertime ice. Thus, the seasonal variability in Northern Hemisphere ice extent varies by only a factor of 2, from a minimum of 7-9 million km² in September to a maximum of 14-16 million km² in March.^{[13][14]}

The ice cover exhibits much greater regional-scale interannual variability than it does hemispherical. For instance, in the region of the Sea of Okhotsk and Japan, maximum ice extent decreased from 1.3 million km² in 1983 to 0.85 million km² in 1984, a decrease of 35%, before rebounding the following year to 1.2 million km².^[13] The regional fluctuations in both hemispheres are such that for any several-year period of the satellite record some regions exhibit decreasing ice coverage while others exhibit increasing ice cover.^[15] The overall trend indicated in the passive microwave record from 1978 through mid-1995 shows that the extent of Arctic sea ice is decreasing 2.7% per decade.^[16] Subsequent work with the satellite passive-microwave data indicates that from late October 1978 through the end of 1996 the extent of Arctic sea ice decreased by 2.9% per decade while the extent of Antarctic sea ice increased by 1.3% per decade.^[17] The Intergovernmental Panel on Climate Change publication *Climate change 2013: The Physical Science Basis* stated that sea ice extent for the Northern Hemisphere showed a decrease of 3.8% ± 0.3% per decade from November 1978 to December 2012.^[18]

Lake ice and river ice

Ice forms on rivers and lakes in response to seasonal cooling. The sizes of the ice bodies involved are too small to exert other than localized climatic effects. However, the freeze-up/break-up processes respond to large-scale and local weather factors, such that considerable interannual variability exists in the dates of appearance and disappearance of the ice. Long series of lake-ice observations can serve as a proxy climate record, and the monitoring of freeze-up and break-up trends may provide a convenient integrated and seasonally specific index of climatic perturbations. Information on river-ice conditions is less useful as a climatic proxy because ice formation is strongly dependent on river-flow regime, which is affected by precipitation, snow melt, and watershed runoff as well as being subject to human interference that directly modifies channel flow, or that indirectly affects the runoff via land-use practices.

Lake freeze-up depends on the heat storage in the lake and therefore on its depth, the rate and temperature of any inflow, and water-air energy fluxes. Information on lake depth is often unavailable, although some indication of the depth of shallow lakes in the Arctic can be obtained from airborne radar imagery during late winter (Sellman *et al.* 1975) and spaceborne optical imagery during summer (Duguay and Lafleur 1997). The timing of breakup is modified by snow depth on the ice as well as by ice thickness and freshwater inflow.

Frozen ground and permafrost

Frozen ground (permafrost and seasonally frozen ground) occupies approximately 54 million km² of the exposed land areas of the Northern Hemisphere (Zhang *et al.*, 2003) and therefore has the largest areal extent of any component of the cryosphere. Permafrost (perennially frozen ground) may occur where mean annual air temperatures (MAAT) are less than -1 or -2 °C and is generally continuous where MAAT are less than -7 °C. In addition, its extent and thickness are affected by ground moisture content, vegetation cover, winter snow depth, and aspect. The global extent of permafrost is still not completely known, but it underlies approximately 20% of Northern Hemisphere land areas. Thicknesses exceed 600 m along the Arctic coast of northeastern Siberia and Alaska, but, toward the margins, permafrost becomes thinner and horizontally discontinuous. The marginal zones will be more immediately subject to any melting caused by a warming trend. Most of the presently existing permafrost formed during previous colder conditions and is therefore relic. However, permafrost may form under present-day polar climates where glaciers retreat or land emergence exposes unfrozen ground. Washburn (1973) concluded that most continuous permafrost is in balance with the present climate at its upper surface, but changes at the base depend on the present climate and geothermal heat flow; in contrast, most discontinuous permafrost is probably unstable or "in such delicate equilibrium that the slightest climatic or surface change will have drastic disequilibrium effects".^[19]

Under warming conditions, the increasing depth of the summer active layer has significant impacts on the hydrologic and geomorphic regimes. Thawing and retreat of permafrost have been reported in the upper Mackenzie Valley and along the southern margin of its occurrence in Manitoba, but such observations are not readily quantified and generalized. Based on average latitudinal gradients of air temperature, an average northward displacement of the southern permafrost boundary by 50-to-150 km could be expected, under equilibrium conditions, for a 1 °C warming.

Only a fraction of the permafrost zone consists of actual ground ice. The remainder (dry permafrost) is simply soil or rock at subfreezing temperatures. The ice volume is generally greatest in the uppermost permafrost layers and mainly comprises pore and segregated ice in Earth material. Measurements of bore-hole temperatures in permafrost can be used as indicators of net changes in temperature regime. Gold and Lachenbruch (1973) infer a 2-4 °C warming over 75 to 100 years at Cape Thompson, Alaska, where the upper 25% of the 400-m thick permafrost is unstable with respect to an equilibrium profile of temperature with depth (for the present mean annual surface temperature of -5 °C). Maritime influences may have biased this estimate, however. At Prudhoe Bay similar data imply a 1.8 °C warming over the last 100 years (Lachenbruch *et al.* 1982). Further complications may be introduced by changes in snow-cover depths and the natural or artificial disturbance of the surface vegetation.

The potential rates of permafrost thawing have been established by Osterkamp (1984) to be two centuries or less for 25-meter-thick permafrost in the discontinuous zone of interior Alaska, assuming warming from -0.4 to 0 °C in 3-4 years, followed by a further 2.6 °C rise. Although the response of permafrost (depth) to temperature change is typically a very slow process (Osterkamp 1984; Koster 1993), there is ample evidence for the fact that the active layer thickness quickly responds to a temperature change (Kane *et al.* 1991). Whether, under a warming or cooling scenario, global climate change will have a significant effect on the duration of frost-free periods in both regions with seasonally and perennially frozen ground.

Glaciers and ice sheets

Ice sheets and glaciers are flowing ice masses that rest on solid land. They are controlled by snow accumulation, surface and basal melt, calving into surrounding oceans or lakes and internal dynamics. The latter results from gravity-driven creep flow ("glacial flow") within the ice body and sliding on the underlying land, which leads to thinning and horizontal spreading.^[20] Any imbalance of this dynamic equilibrium between mass gain, loss and transport due to flow results in either growing or shrinking ice bodies.

Ice sheets are the greatest potential source of global freshwater, holding approximately 77% of the global total. This corresponds to 80 m of world sea-level equivalent, with Antarctica accounting for 90% of this. Greenland accounts for most of the remaining 10%, with other ice bodies and glaciers accounting for less than 0.5%. Because of their size in relation to annual rates of snow accumulation and melt, the residence time of water in ice sheets can extend to 100,000 or 1 million years. Consequently, any climatic perturbations produce slow responses, occurring over glacial and interglacial periods. Valley glaciers respond rapidly to climatic fluctuations with typical response times of 10-50 years.^[21] However, the response of individual glaciers may be asynchronous to the same climatic forcing because of differences in glacier length, elevation, slope, and speed of motion. Oerlemans (1994) provided evidence of coherent global glacier retreat which could be explained by a linear warming trend of 0.66 °C per 100 years.^[21]

While glacier variations are likely to have minimal effects upon global climate, their recession may have contributed one third to one half of the observed 20th Century rise in sea level (Meier 1984; IPCC 1996). Furthermore, it is extremely likely that such extensive glacier recession as is currently observed in the Western Cordillera of North America,^[22] where runoff from glacierized basins is used for irrigation and hydropower, involves significant hydrological and ecosystem impacts. Effective water-resource planning and impact mitigation in such areas depends upon developing a sophisticated knowledge of the status of glacier ice and the mechanisms that cause it to change. Furthermore, a clear understanding of the mechanisms at work is crucial to interpreting the global-change signals that are contained in the time series of glacier mass balance records.

Combined glacier mass balance estimates of the large ice sheets carry an uncertainty of about 20%. Studies based on estimated snowfall and mass output tend to indicate that the ice sheets are near balance or taking some water out of the oceans.^[23] Marinebased studies^[24] suggest sea-level rise from the Antarctic or rapid ice-shelf basal melting. Some authors (Paterson 1993; Alley 1997) have suggested that the difference between the observed rate of sea-level rise (roughly 2 mm/y) and the explained rate of sea-level rise from melting of mountain glaciers, thermal expansion of the ocean, etc. (roughly 1 mm/y or less) is similar to the modeled imbalance in the Antarctic (roughly 1 mm/y of sea-level rise; Huybrechts 1990), suggesting a contribution of sea-level rise from the Antarctic.

Relationships between global climate and changes in ice extent are complex. The mass balance of land-based glaciers and ice sheets is determined by the accumulation of snow, mostly in winter, and warm-season ablation due primarily to net radiation and turbulent heat fluxes to melting ice and snow from warm-air advection,^{[25][26]}(Munro 1990). However, most of Antarctica never experiences surface melting.^[27] Where ice masses terminate in the ocean, iceberg calving is the major contributor to mass loss. In this situation, the ice margin may extend out into deep water as a floating ice shelf, such as that in the Ross Sea. Despite the possibility that global warming could result in losses to the Greenland ice sheet being offset by gains to the Antarctic ice sheet,^[28] there is major concern about the possibility of a West Antarctic Ice Sheet collapse. The West Antarctic Ice Sheet is grounded on bedrock below sea level, and its collapse has the potential of raising the world sea level 6–7 m over a few hundred years.

Most of the discharge of the West Antarctic Ice Sheet is via the five major ice streams (faster flowing ice) entering the Ross Ice Shelf, the Rutford Ice Stream entering Ronne-Filchner shelf of the Weddell Sea, and the Thwaites Glacier and Pine Island Glacier entering the Amundsen Ice Shelf. Opinions differ as to the present mass balance of these systems (Bentley 1983, 1985), principally because of the limited data. The West Antarctic Ice Sheet is stable so long as the Ross Ice Shelf is constrained by drag along its lateral boundaries and pinned by local grounding.

See also

- Active layer
- Cryobiology
- Glaciers
- Hydrosphere
- Ice
- Ice cap
- Ice sheet
- Permafrost
- Sea ice
- Snow

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- International Glaciological Society (IGS)
- Global Outlook for Ice and Snow Assessment on the state and future of the Cryosphere, by the UN Environment Programme, June 2007
- Cryosphere overview map, from the UN Environment Programme
- Department of Atmospheric Sciences at the University of Illinois: Cryosphere Today
- Canadian Cryospheric Information Network
- State of the Canadian Cryosphere
- Near-real-time overview of global ice concentration and snow extent
- National Snow and Ice Data Center
- ResearchChannel – Cryospheric Response to Climate Change. A video produced by the University of Washington, March 2008

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Categories: Arctic geography terminology | Earth sciences | Environment of Antarctica | Forms of water | Hydrology | Structure of the Earth

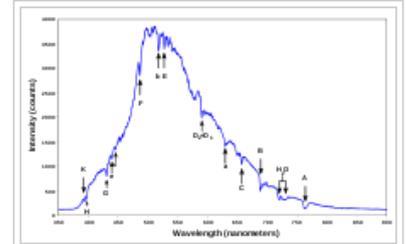
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Diffuse sky radiation

From Wikipedia, the free encyclopedia

Diffuse sky radiation is solar radiation reaching the Earth's surface after having been scattered from the direct solar beam by molecules or suspensoids in the atmosphere. It is also called **skylight**, **diffuse skylight**, or **sky radiation** and is the reason for changes in the color of the sky. Of the total light removed from the direct solar beam by scattering in the atmosphere (approximately 25% of the incident radiation when the sun is high in the sky, depending on the amount of dust and haze in the atmosphere), about two-thirds ultimately reaches the earth as diffuse sky radiation. When the sun is at the zenith in a cloudless sky, with 1361 W/m^2 ^[1] above the atmosphere, direct sunlight is about 1050 W/m^2 , and total insolation about 1120 W/m^2 .^[2] This implies that under these conditions the diffuse radiation is only about 70 W/m^2 out of the original 1361 W/m^2 .

The dominant radiative scattering processes in the atmosphere (Rayleigh scattering and Mie scattering) are elastic in nature, by which light can be deviated from its path without being absorbed and with no change in wavelength.



A spectrum taken of blue sky clearly showing solar Fraunhofer lines and atmospheric water absorption band.

Contents

- 1 Color
- 2 Neutral points
- 3 Under an overcast sky
- 4 As a part of total radiation
- 5 See also
- 6 References
- 7 Further reading
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Color

The sunlit sky is blue because air scatters short-wavelength light more than longer wavelengths. Since blue light is at the short-wavelength end of the visible spectrum, it is more strongly scattered in the atmosphere than long-wavelength red light. The result is that when looking toward parts of the sky other than the sun, human eye perceives them to be blue.^[3] The color perceived is similar to that obtained by a monochromatic blue of a wavelength of 474–476 nm mixed with white light, *i.e.*, an unsaturated blue light.^[4] The explanation of the blue color by Rayleigh in 1871 is one of the most famous examples of the application of dimensional analysis in solving a problem in physics.^[5]

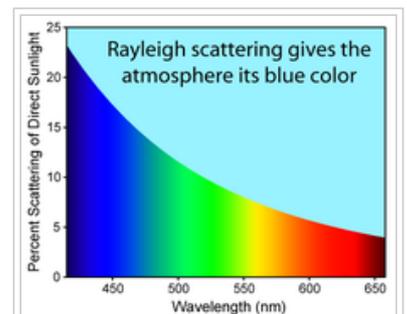
Near sunrise and sunset, most of the sunlight arrives nearly tangentially to the Earth's surface; thus, the light's path through the atmosphere is so long that much of the blue and even green light is scattered out along the way, leaving the sun rays and the clouds it illuminates red. Therefore, when looking at the sunset and sunrise, we see the colour red more than the other colors.

Scattering and absorption are major causes of the attenuation of radiation by the atmosphere. Scattering varies as a function of the ratio of the particle diameter to the wavelength of the radiation. When this ratio is less than about one-tenth, Rayleigh scattering occurs. In this case the scattering coefficient varies inversely with the fourth power of the wavelength. At larger ratios, scattering varies in a complex fashion, described for spherical particles by the Mie theory; the laws of geometric optics begin to apply at a ratio of the order of 10.

In the example of the sky at the zenith, it is blue during broad daylight due to Rayleigh scattering involving diatomic gases (N_2 , O_2). Near sunset and especially during twilight, ozone (O_3) absorption significantly contributes to maintaining the sky's blue color.



Clear blue sky



Comparison of the scattering efficiency of blue light in the atmosphere compared to red light.

Neutral points

There are four commonly detectable points of zero polarization of diffuse sky radiation (known as neutral points) lying along the vertical circle through the sun.

- The **Arago point**, named after its discoverer, is customarily located at about 20° above the antisolar point; but it lies at higher altitudes in turbid air. The latter property makes the Arago distance a useful measure of atmospheric turbidity.
- The **Babinet point**, discovered by Jacques Babinet in 1840, is located about 15° to 20° above the sun, hence it is difficult to observe because of solar glare.
- The **Brewster point**, discovered by David Brewster in 1840, is located about 15° to 20° below the sun; hence it is difficult to observe because of solar glare.
- The fourth point, located at about 20° below the antisolar point, visible only at higher altitudes in the air or in space.^[6]

Under an overcast sky

There is essentially no direct sunlight under an overcast sky, so all light is then diffuse sky radiation. The flux of light is not very wavelength dependent because the cloud droplets are larger than the light's wavelength and scatter all colors approximately equally. The light passes through the translucent clouds in a manner similar to frosted glass. The intensity ranges (roughly) from $\frac{1}{6}$ of direct sunlight for relatively thin clouds down to $\frac{1}{1000}$ of direct sunlight under the extreme of thickest storm clouds.

As a part of total radiation

One of the equations for total solar radiation is:^[7]

$$H_t = H_b R_b + H_d R_d + (H_b + H_d) R_r$$

where H_b is the beam radiation irradiance, R_b is the tilt factor for beam radiation, H_d is the diffuse radiation irradiance, R_d is the tilt factor for diffuse radiation and R_r is the tilt factor for reflected radiation.

R_b is given by:

$$R_b = \frac{\sin(\delta) \sin(\phi - \beta) + \cos(\delta) \cos(h) \cos(\phi - \beta)}{\sin(\delta) \sin(\phi) + \cos(\delta) \cos(h) \cos(\phi)}$$

where δ is the solar declination, Φ is the latitude, β is an angle from the horizontal and h is the solar hour angle.

R_d is given by:

$$R_d = \frac{1 + \cos(\beta)}{2}$$

and R_r by:

$$R_r = \frac{\rho(1 - \cos(\beta))}{2}$$

where ρ is the reflectivity of the surface.

See also

- Atmospheric diffraction
- Aerial perspective
- Cyanometer
- Daylight
- Nighttime airglow
- Tyndall effect
- Rayleigh scattering
- Rayleigh Sky Model

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External links

- Dr. C. V. Raman lecture: Why is the sky blue?
- Why is the sky blue?
- Blue Sky and Rayleigh Scattering
- Atmospheric Optics (.pdf), Dr. Craig Bohren

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Categories: Sun | Light | Visibility | Atmospheric optical phenomena

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Extinction (astronomy)

From Wikipedia, the free encyclopedia

In astronomy, **extinction** is the absorption and scattering of electromagnetic radiation by dust and gas between an emitting astronomical object and the observer. Interstellar extinction was first documented as such in 1930 by Robert Julius Trumpler.^{[1][2]} However, its effects had been noted in 1847 by Friedrich Georg Wilhelm von Struve,^[3] and its effect on the colors of stars had been observed by a number of individuals who did not connect it with the general presence of galactic dust. For stars that lie near the plane of the Milky Way and are within a few thousand parsecs of the Earth, extinction in the visual band of frequencies (photometric system) is on the order of 1.8 magnitudes per kiloparsec.^[4]

For Earth-bound observers, extinction arises both from the interstellar medium (ISM) and the Earth's atmosphere; it may also arise from circumstellar dust around an observed object. The strong atmospheric extinction in some wavelength regions (such as X-ray, ultraviolet, and infrared) requires the use of space-based observatories. Since blue light is much more strongly attenuated than red light, extinction causes objects to appear redder than expected, a phenomenon referred to as **interstellar reddening**.^[5]

Contents

- 1 Interstellar reddening
- 2 General characteristics
- 3 Measuring extinction towards an object
- 4 The 2175-angstrom feature
- 5 Extinction curves of other galaxies
- 6 Atmospheric extinction
- 7 References
- 8 Further reading

Interstellar reddening

In astronomy, interstellar reddening is a phenomenon associated with interstellar extinction where the spectrum of electromagnetic radiation from a radiation source changes characteristics from that which the object originally emitted. Reddening occurs due to the light scattering off dust and other matter in the interstellar medium. Interstellar reddening is a different phenomenon from redshift, which is the proportional frequency shifts of spectra without distortion. Reddening preferentially removes shorter wavelength photons from a radiated spectrum while leaving behind the longer wavelength photons (in the optical, light that is redder), leaving the spectroscopic lines unchanged.

In any photometric system interstellar reddening can be described by color excess, defined as the difference between an object's observed color index and its intrinsic color index (sometimes referred to as its normal color index). An object's intrinsic color index is the theoretical color index which it would have if unaffected by extinction. In the UBV photometric system the color excess E_{B-V} is related to the B-V colour by:

$$E_{B-V} = (B - V)_{\text{observed}} - (B - V)_{\text{intrinsic}}$$

General characteristics

Interstellar reddening occurs because interstellar dust absorbs and scatters blue light waves more than red light waves, making stars appear redder than they are. This is similar to the effect seen when dust particles in the atmosphere of Earth contribute to red sunsets.^[6]

Broadly speaking, interstellar extinction is strongest at short wavelengths, generally observed by using techniques from spectroscopy. Extinction results in a change in the shape of an observed spectrum. Superimposed on this general shape are absorption features (wavelength bands where the intensity is lowered) that have a variety of origins and can give clues as to the chemical composition of the interstellar material, e.g. dust grains. Known absorption features include the 2175 Å bump, the diffuse interstellar bands, the 3.1 μm water ice feature, and the 10 and 18 μm silicate features.

In the solar neighborhood, the rate of interstellar extinction in the Johnson-Cousins V-band is usually taken to be 0.7-1.0 mag/kpc—simply an average due to the *clumpiness* of interstellar dust.^{[7][8][9]} In general, however, this means that a star will have its brightness reduced by about a factor of 2 in the V-band for every kiloparsec it is farther away from us.

The amount of extinction can be significantly higher than this in specific directions. For example, some regions of the Galactic Center have more than 30 magnitudes of extinction in the optical, meaning that less than 1 optical photon in 10^{12} passes through.^[10] This results in the so-called zone of avoidance, where our view of the extra-galactic sky is severely hampered, and background galaxies, such as Dwingeloo 1, were only discovered recently through observations in radio and infrared.

The general shape of the ultraviolet through near-infrared (0.125 to 3.5 μm) extinction curve in our own galaxy, the Milky Way, is fairly well characterized by the single parameter $R(V)$ (which is different along different lines of sight through the galaxy),^{[11][12]} but there are known deviations from this single parameter characterization.^[13] Extending the extinction law into the mid-infrared wavelength range is difficult due to the lack of suitable targets and various contributions by absorption features.^[14]

$R(V)$ is defined to be $A(V)/E(B-V)$, and measures the total, $A(V)$, to selective, $E(B-V) = A(B) - A(V)$, extinction in set bands. $A(B)$ and $A(V)$ are the **total extinction** at the B and V filter bands. Another measure used in the literature is the **absolute extinction** $A(\lambda)/A(V)$ at wavelength λ , comparing the total extinction at that wavelength to that at the V band.

$R(V)$ is known to be correlated with the average size of the dust grains causing the extinction. For our own galaxy, the Milky Way, the typical value for $R(V)$ is 3.1,^[15] but is found to be between 2.5 and 6 for different lines of sight.

The relationship between the total extinction, $A(V)$ (measured in magnitudes), and the column density of neutral hydrogen atoms column, N_H (usually measured in cm^{-2}), shows how the gas and dust in the interstellar medium are related. From studies using ultraviolet spectroscopy of reddened stars and X-ray scattering halos in the Milky Way, Predehl and Schmitt^[16] found the relationship between N_H and $A(V)$ to be approximately:

$$\frac{N_H}{A(V)} \approx 1.8 \times 10^{21} \text{ atoms cm}^{-2} \text{ mag}^{-1}$$

(see also:^{[17][18][19]}).

Astronomers have determined the three-dimensional distribution of extinction in the solar circle of our galaxy, using visible and near-infrared stellar observations and a model of the distribution of stars in the galaxy.^{[20][21]} The dust giving rise to the extinction lies along the spiral arms, as observed in other spiral galaxies.

Measuring extinction towards an object

To measure the extinction curve for a star, the star's spectrum is compared to the observed spectrum of a similar star known not to be affected by extinction (unreddened).^[22] It is also possible to use a theoretical spectrum instead of the observed spectrum for the comparison, but this is less common. In the case of emission nebulae, it is common to look at the ratio of two emission lines which should not be affected by the temperature and density in the nebula. For example, the ratio of hydrogen alpha to hydrogen beta emission is always around 2.85 under a wide range of conditions prevailing in nebulae. A ratio other than 2.85 must therefore be due to extinction, and the amount of extinction can thus be calculated.

The 2175-angstrom feature

One prominent feature in measured extinction curves of many objects within the Milky Way is a broad 'bump' at about 2175 Å, well into the ultraviolet region of the electromagnetic spectrum. This feature was first observed in the 1960s,^{[23][24]} but its origin is still not well understood. Several models have been presented to account for this bump which include graphitic grains with a mixture of PAH molecules. Investigations of interstellar grains embedded in interplanetary dust particles (IDP) observed this feature and identified the carrier with organic carbon and amorphous silicates present in the grains.^[25]

Extinction curves of other galaxies

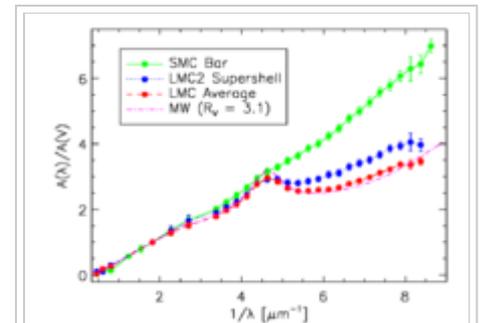
The form of the standard extinction curve depends on the composition of the ISM, which varies from galaxy to galaxy. In the Local Group, the best-determined extinction curves are those of the Milky Way, the Small Magellanic Cloud (SMC) and the Large Magellanic Cloud (LMC).

In the LMC, there is significant variation in the characteristics of the ultraviolet extinction with a weaker 2175 Å bump and stronger far-UV extinction in the region associated with the LMC2 supershell (near the 30 Doradus starbursting region) than seen elsewhere in the LMC and in the Milky Way.^{[27][28]} In the SMC, more extreme variation is seen with no 2175 Å and very strong far-UV extinction in the star forming Bar and fairly normal ultraviolet extinction seen in the more quiescent Wing.^{[29][30][31]}

This gives clues as to the composition of the ISM in the various galaxies. Previously, the different average extinction curves in the Milky Way, LMC, and SMC were thought to be the result of the different metallicities of the three galaxies: the LMC's metallicity is about 40% of that of the Milky Way, while the SMC's is about 10%. Finding extinction curves in both the LMC and SMC which are similar to those found in the Milky Way^[26] and finding extinction curves in the Milky Way that look more like those found in the LMC2 supershell of the LMC^[32] and in the SMC Bar^[33] has given rise to a new interpretation. The variations in the curves seen in the Magellanic Clouds and Milky Way may instead be caused by processing of the dust grains by nearby star formation. This interpretation is supported by work in starburst galaxies (which are undergoing intense star formation episodes) that their dust lacks the 2175 Å bump.^{[34][35]}

Atmospheric extinction

Atmospheric extinction gives the rising or setting Sun an orange hue and varies with location and altitude. Astronomical observatories generally are able to characterise the local extinction curve very accurately, to allow observations to be corrected for the effect. Nevertheless, the atmosphere is completely opaque to many wavelengths requiring the use of satellites to make observations.



Plot showing the average extinction curves for the MW, LMC2, LMC, and SMC Bar.^[26] The curves are plotted versus 1/wavelength to emphasize the UV.

Atmospheric extinction has three main components: Rayleigh scattering by air molecules, light scattering by aerosols, and molecular absorption. Molecular absorption is often referred to as telluric absorption, as it is caused by the Earth (*telluric* is a synonym for *terrestrial*). The most important sources of telluric absorption are molecular oxygen and ozone, which absorb strongly in the near-ultraviolet, and water, which absorbs strongly in the infrared.

The amount of atmospheric extinction depends on the altitude of an object, being lowest at the zenith and at a maximum near the horizon. It is calculated by multiplying the standard atmospheric extinction curve by the mean airmass calculated over the duration of the observation.

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Categories: Observational astronomy | Galactic astronomy | Extragalactic astronomy

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Fossil fuel

From Wikipedia, the free encyclopedia

A **fossil fuel** is a fuel formed by natural processes, such as anaerobic decomposition of buried dead organisms, containing energy originating in ancient photosynthesis.^[1] The age of the organisms and their resulting fossil fuels is typically millions of years, and sometimes exceeds 650 million years.^[2] Fossil fuels contain high percentages of carbon and include petroleum, coal, and natural gas.^[3] Other commonly used derivatives include kerosene and propane. Fossil fuels range from volatile materials with low carbon to hydrogen ratios like methane, to liquids like petroleum, to nonvolatile materials composed of almost pure carbon, like anthracite coal. Methane can be found in hydrocarbon fields either alone, associated with oil, or in the form of methane clathrates.

The theory that fossil fuels formed from the fossilized remains of dead plants^[4] by exposure to heat and pressure in the Earth's crust over millions of years^[5] was first introduced by Georgius Agricola in 1556 and later by Mikhail Lomonosov in the 18th century.

The Energy Information Administration estimates that in 2007 the primary sources of energy consisted of petroleum 36.0%, coal 27.4%, natural gas 23.0%, amounting to an 86.4% share for fossil fuels in primary energy consumption in the world.^[6] Non-fossil sources in 2006 included nuclear 8.5%, hydroelectric 6.3%, and others (geothermal, solar, tidal, wind, wood, waste) amounting to 0.9%.^[7] World energy consumption was growing about 2.3% per year.

Although fossil fuels are continually being formed via natural processes, they are generally considered to be non-renewable resources because they take millions of years to form and the known viable reserves are being depleted much faster than new ones are being made.^{[8][9]}

The use of fossil fuels raises serious environmental concerns. The burning of fossil fuels produces around 21.3 billion tonnes (21.3 gigatonnes) of carbon dioxide (CO₂) per year. It is estimated that natural processes can only absorb about half of that amount, so there is a net increase of 10.65 billion tonnes of atmospheric carbon dioxide per year.^[10] Carbon dioxide is a greenhouse gas that increases radiative forcing and contributes to global warming. A global movement towards the generation of renewable energy is underway to help reduce global greenhouse gas emissions.



Coal, one of the fossil fuels

Contents

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Origin

Petroleum and natural gas are formed by the anaerobic decomposition of remains of organisms including phytoplankton and zooplankton that settled to the sea (or lake) bottom in large quantities under anoxic conditions, millions of years ago. Over geological time, this organic matter, mixed with mud, got buried under heavy layers of sediment. The resulting high levels of heat and pressure caused the organic matter to chemically alter, first into a waxy material known as kerogen which is found in oil shales, and then with more heat into liquid and gaseous hydrocarbons in a process known as catagenesis. Despite these heat driven transformations (which may increase the energy density compared to typical organic matter), the embedded energy is still photosynthetic in origin.^[1]

There is a wide range of organic, or hydrocarbon, compounds in any given fuel mixture. The specific mixture of hydrocarbons gives a fuel its characteristic properties, such as boiling point, melting point, density, viscosity, etc. Some fuels like natural gas, for instance, contain only very low boiling, gaseous components. Others such as gasoline or diesel contain much higher boiling components.

Terrestrial plants, on the other hand, tend to form coal and methane. Many of the coal fields date to the Carboniferous period of Earth's history. Terrestrial plants also form type III kerogen, a source of natural gas.

Importance

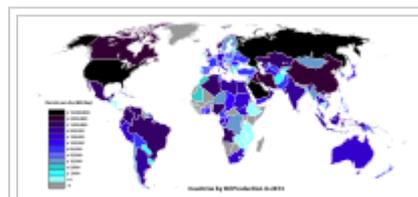
Fossil fuels are of great importance because they can be burned (oxidized to carbon dioxide and water), producing significant amounts of energy per unit mass. The use of coal as a fuel predates recorded history. Coal was used to run furnaces for the melting of metal ore. Semi-solid hydrocarbons from seeps were also burned in ancient times,^[12] but these materials were mostly used for waterproofing and embalming.^[13]

Commercial exploitation of petroleum, largely as a replacement for oils from animal sources (notably whale oil), for use in oil lamps began in the 19th century.^[14]

Natural gas, once flared-off as an unneeded byproduct of petroleum production, is now considered a very valuable resource.^[15] Natural gas deposits are also the main source of the element helium.

Heavy crude oil, which is much more viscous than conventional crude oil, and tar sands, where bitumen is found mixed with sand and clay, are becoming more important as sources of fossil fuel.^[16] Oil shale and similar materials are sedimentary rocks containing kerogen, a complex mixture of high-molecular weight organic compounds, which yield synthetic crude oil when heated (pyrolyzed). These materials have yet to be exploited commercially.^[17] These fuels can be employed in internal combustion engines, fossil fuel power stations and other uses.

Prior to the latter half of the 18th century, windmills and watermills provided the energy needed for industry such as milling flour, sawing wood or pumping water, and burning wood or peat provided domestic heat. The widescale use of fossil fuels, coal at first and petroleum later, to fire steam engines enabled the Industrial



Since oil fields are located only at certain places on earth,^[11] only a select group of countries are oil-independent; the other countries depend on the oil-production capacities of these countries



A petrochemical refinery in Grangemouth, Scotland, UK

Revolution. At the same time, gas lights using natural gas or coal gas were coming into wide use. The invention of the internal combustion engine and its use in automobiles and trucks greatly increased the demand for gasoline and diesel oil, both made from fossil fuels. Other forms of transportation, railways and aircraft, also required fossil fuels. The other major use for fossil fuels is in generating electricity and as feedstock for the petrochemical industry. Tar, a leftover of petroleum extraction, is used in construction of roads.

Reserves

Levels of primary energy sources are the reserves in the ground. Flows are production of fossil fuels from these reserves. The most important part of primary energy sources are the carbon based fossil energy sources. Coal, oil, and natural gas provided 79.6% of primary energy production during 2002 (in million tonnes of oil equivalent (mtoe)) (34.9+23.5+21.2).

Levels (proved reserves) during 2005–2006

- Coal: 997,748 million short tonnes (905 billion metric tonnes),^[18] 4,416 billion barrels (702.1 km³) of oil equivalent
- Oil: 1,119 billion barrels (177.9 km³) to 1,317 billion barrels (209.4 km³)^[19]
- Natural gas: 6,183–6,381 trillion cubic feet (175–181 trillion cubic metres),^[19] 1,161 billion barrels (184.6 × 10⁹ m³) of oil equivalent

Flows (daily production) during 2006

- Coal: 18,476,127 short tonnes (16,761,260 metric tonnes),^[20] 52,000,000 barrels (8,300,000 m³) of oil equivalent per day
- Oil: 84,000,000 barrels per day (13,400,000 m³/d)^[21]
- Natural gas: 104,435 billion cubic feet (2,963 billion cubic metres),^[22] 19,000,000 barrels (3,000,000 m³) of oil equivalent per day

Limits and alternatives

P. E. Hodgson, a Senior Research Fellow Emeritus in Physics at Corpus Christi College, Oxford, expects the world energy use is doubling every fourteen years and the need is increasing faster still and he insisted in 2008 that the world oil production, a main resource of fossil fuel, is expected to peak in ten years and thereafter fall.^[23]

The principle of supply and demand holds that as hydrocarbon supplies diminish, prices will rise. Therefore, higher prices will lead to increased alternative, renewable energy supplies as previously uneconomic sources become sufficiently economical to exploit. Artificial gasolines and other renewable energy sources currently require more expensive production and processing technologies than conventional petroleum reserves, but may become economically viable in the near future. Different alternative sources of energy include nuclear, hydroelectric, solar, wind, and geothermal.

One of the more promising energy alternatives is the use of inedible feed stocks and biomass for carbon dioxide capture as well as biofuel. While these processes are not without problems, they are currently in practice around the world. Biodiesels are being produced by several companies and source of great research at several universities. Some of the most common and promising processes of conversion of renewable lipids into usable fuels is through hydrotreating and decarboxylation.

Environmental effects



An oil well in the Gulf of Mexico

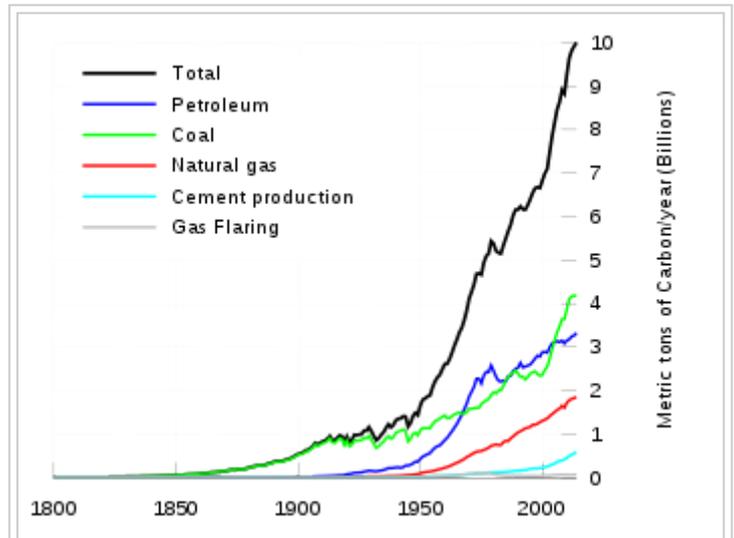
The United States holds less than 5% of the world's population, but due to large houses and private cars, uses more than a quarter of the world's supply of fossil fuels.^[24] In the United States, more than 90% of greenhouse gas emissions come from the combustion of fossil fuels.^[25] Combustion of fossil fuels also produces other air pollutants, such as nitrogen oxides, sulfur dioxide, volatile organic compounds and heavy metals.

According to Environment Canada:

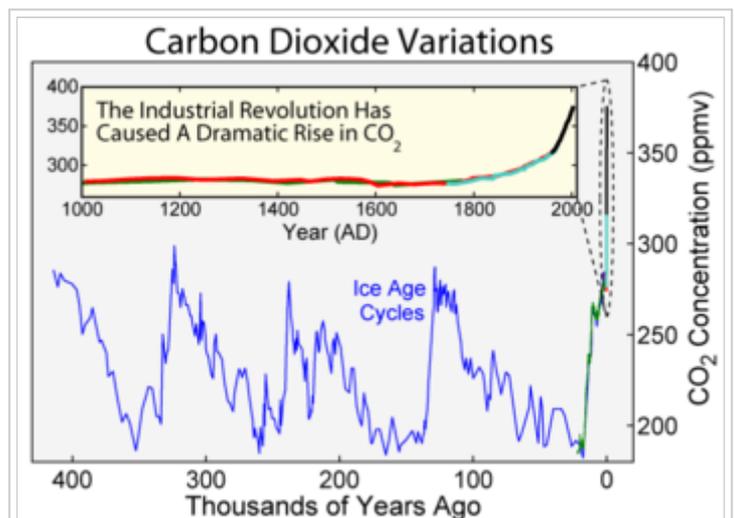
"The electricity sector is unique among industrial sectors in its very large contribution to emissions associated with nearly all air issues. Electricity generation produces a large share of Canadian nitrogen oxides and sulphur dioxide emissions, which contribute to smog and acid rain and the formation of fine particulate matter. It is the largest uncontrolled industrial source of mercury emissions in Canada. Fossil fuel-fired electric power plants also emit carbon dioxide, which may contribute to climate change. In addition, the sector has significant impacts on water and habitat and species. In particular, hydropower dams and transmission lines have significant effects on water and biodiversity."^[26]

According to U.S. Scientist Jerry Mahlman and USA Today: Mahlman, who crafted the IPCC language used to define levels of scientific certainty, says the new report will lay the blame at the feet of fossil fuels with "virtual certainty," meaning 99% sure. That's a significant jump from "likely," or 66% sure, in the group's last report in 2001, Mahlman says. His role in this year's effort involved spending two months reviewing the more than 1,600 pages of research that went into the new assessment.^[27]

Combustion of fossil fuels generates sulfuric, carbonic, and nitric acids, which fall to Earth as acid rain, impacting both natural areas and the built environment. Monuments and sculptures made from marble and limestone are particularly vulnerable, as the acids dissolve calcium carbonate.



Global fossil carbon emission by fuel type, 1800–2007. Note: Carbon only represents 27% of the mass of CO₂



Carbon dioxide variations over the last 400,000 years, showing a rise since the industrial revolution

Fossil fuels also contain radioactive materials, mainly uranium and thorium, which are released into the atmosphere. In 2000, about 12,000 tonnes of thorium and 5,000 tonnes of uranium were released worldwide from burning coal.^[28] It is estimated that during 1982, US coal burning released 155 times as much radioactivity into the atmosphere as the Three Mile Island accident.^[29]

Burning coal also generates large amounts of bottom ash and fly ash. These materials are used in a wide variety of applications, utilizing, for example, about 40% of the US production.^[30]

Harvesting, processing, and distributing fossil fuels can also create environmental concerns. Coal mining methods, particularly mountaintop removal and strip mining, have negative environmental impacts, and offshore oil drilling poses a hazard to aquatic organisms. Oil refineries also have negative environmental impacts, including air and water pollution. Transportation of coal requires the use of diesel-powered locomotives, while crude oil is typically transported by tanker ships, each of which requires the combustion of additional fossil fuels.

Environmental regulation uses a variety of approaches to limit these emissions, such as command-and-control (which mandates the amount of pollution or the technology used), economic incentives, or voluntary programs.

An example of such regulation in the USA is the "EPA is implementing policies to reduce airborne mercury emissions. Under regulations issued in 2005, coal-fired power plants will need to reduce their emissions by 70 percent by 2018."^[31]

In economic terms, pollution from fossil fuels is regarded as a negative externality. Taxation is considered one way to make societal costs explicit, in order to 'internalize' the cost of pollution. This aims to make fossil fuels more expensive, thereby reducing their use and the amount of pollution associated with them, along with raising the funds necessary to counteract these factors.

According to Rodman D. Griffin, "The burning of coal and oil have saved inestimable amounts of time and labor while substantially raising living standards around the world".^[32] Although the use of fossil fuels may seem beneficial to our lives, this act is playing a role on global warming and it is said to be dangerous for the future.^[32]

Moreover, these environmental pollutions impacts on the human beings because its particles of the fossil fuel on the air cause negative health effects when inhaled by people. These health effects include premature death, acute respiratory illness, aggravated asthma, chronic bronchitis and decreased lung function. So, the poor, undernourished, very young and very old, and people with preexisting respiratory disease and other ill health, are more at risk.^[33]

Industry

Economic effects

Europe spent €406 billion on importing fossil fuels in 2011 and €545 billion in 2012. This is around three times more than the cost of the Greek bailout up to 2013. In 2012 wind energy in Europe avoided €9.6 billion of fossil fuel costs.^[34] A 2014 report by the International Energy Agency said that the fossil fuels industry collects \$550 billion a year in global government subsidies.^[35] This amount was \$490 billion in 2014, but would have been \$610 billion without agreements made in 2009.^[36]

A 2015 report studied 20 fossil fuel companies and found that, while highly profitable, the hidden economic cost to society was also large.^{[37][38]} The report spans the period 2008–2012 and notes that: "For all companies and all years, the economic cost to society of their CO₂ emissions was greater than their after-tax profit, with the single exception of ExxonMobil in 2008."^{[37]:4} Pure coal companies fare even worse: "the economic cost to

society exceeds total revenue in all years, with this cost varying between nearly \$2 and nearly \$9 per \$1 of revenue."^{[37]:5} In this case, total revenue includes "employment, taxes, supply purchases, and indirect employment."^{[37]:4}

See also

- Abiogenic petroleum origin proposes that petroleum is not a fossil fuel
- Bioremediation
- Carbon bubble
- Environmental impact of the energy industry
- Fossil Fools Day
- Fossil fuel divestment
- Fossil fuel drilling
- Fossil fuel exporters
- Fossil fuel phase-out
- Fossil fuels lobby
- Hydraulic fracturing
- Liquefied petroleum gas
- Low-carbon power
- Peak coal
- Peak gas
- Petroleum industry
- Shale gas

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External links

- "The Coming Energy Crisis?" – essay by James L. Williams of WTRG Economics and A. F. Alhajji of Ohio Northern University
- "Powering the Future" – Michael Parfit (*National Geographic*)
- "Federal Fossil Fuel Subsidies and Greenhouse Gas Emissions"
- Fossil Fuel Subsidies in Europe
- Oil companies hit by 'state' cyber attacks



Wikiquote has quotations related to: ***Fossil fuel***

Debate

- The Origin of Methane (and Oil) in the Crust of the Earth – Thomas Gold (Internet Archives)
- Abiotic theory debunked "fossil fuel" myth

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Categories: Fossil fuels

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Geosphere

From Wikipedia, the free encyclopedia

There are several conflicting definitions for **geosphere**.

The **geosphere** may be taken as the collective name for the lithosphere, the hydrosphere, the cryosphere, and the atmosphere.^[1]

シムダ
入ル
水
維
田 Look up *geosphere* in Wiktionary, the free dictionary.

In Aristotelian physics, the term was applied to four spherical *natural places*, concentrically nested around the center of the Earth, as described in the lectures *Physica* and *Meteorologica*. They were believed to explain the motions of the four *terrestrial elements*: *Earth*, *Water*, *Air* and *Fire*.

In modern texts and in Earth system science, geosphere refers to the solid parts of the Earth; it is used along with atmosphere, hydrosphere, and biosphere to describe the systems of the Earth (the interaction of these systems with the magnetosphere is sometimes listed). In that context, sometimes the term lithosphere is used instead of geosphere or solid Earth. The lithosphere, however, only refers to the uppermost layers of the solid Earth (oceanic and continental crustal rocks and uppermost mantle).^[2]

Since space exploration began, it has been observed that the extent of the ionosphere or plasmasphere is highly variable, and often much larger than previously appreciated, at times extending to the boundaries of the Earth's magnetosphere or geomagnetosphere.^[3] This highly variable outer boundary of *geogenic* matter has been referred to as the "geopause,"^[4] to suggest the relative scarcity of such matter beyond it, where the solar wind dominates.

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Great Oxygenation Event

From Wikipedia, the free encyclopedia

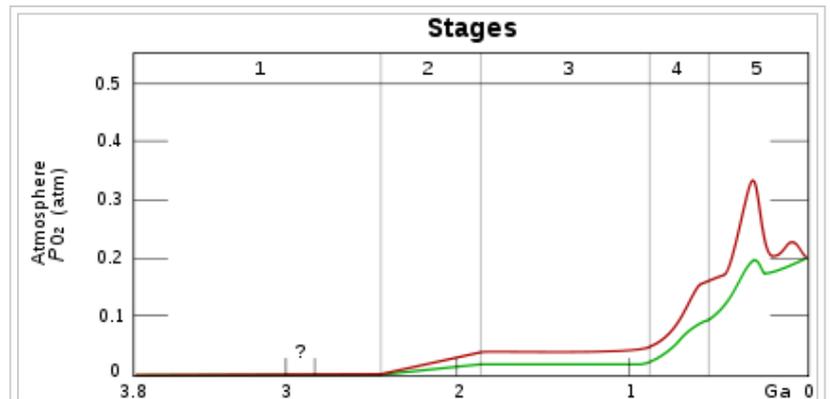
The **Great Oxygenation Event** (GOE, also called the **Oxygen Catastrophe**, **Oxygen Crisis**, **Oxygen Holocaust**, **Oxygen Revolution**, or **Great Oxidation**) was the biologically induced appearance of dioxygen (O₂) in Earth's atmosphere.^[2] Although geological, isotopic, and chemical evidence suggest that this major environmental change happened around 2.45 billion years ago (2.45 Ga),^[3] (during the Siderian period) the actual causes and the exact date of the event are not clear.^[4] The current geochemical and biomarker evidence for the development of oxygenic photosynthesis before the Great Oxidation Event has been mostly inconclusive.^[5]

Oceanic cyanobacteria, which evolved into multicellular forms more than 2.3 billion years ago (approximately 200 million years before the GOE),^[6] are believed to have become the first microbes to produce oxygen by photosynthesis.^[7] Before the GOE, any free oxygen they produced was chemically captured by dissolved iron or organic matter. The GOE was the point in time when these oxygen sinks became saturated, at which point oxygen, produced by the cyanobacteria, was free to escape into the atmosphere.

The increased production of oxygen set Earth's original atmosphere off balance.^[8] Free oxygen is toxic to obligate anaerobic organisms, and the rising concentrations may have destroyed most such organisms at the time. Cyanobacteria were therefore responsible for one of the most significant extinction events in Earth's history. Besides marine cyanobacteria, there is also evidence of cyanobacteria on land.

A spike in chromium contained in ancient rock deposits formed underwater shows they had accumulated chromium washed off from continental shelves. Chromium is not easily dissolved and its release from rocks would have required the presence of a powerful acid. One such acid, sulfuric acid, might have been created through bacterial reactions with pyrite.^[9] Mats of oxygen-producing cyanobacteria can produce a thin layer, one or two millimeters thick, of oxygenated water in an otherwise anoxic environment even under thick ice, and before oxygen started accumulating in the atmosphere, these organisms would already be adapted to oxygen.^[10] Additionally, the free oxygen would have reacted with atmospheric methane, a greenhouse gas, greatly reducing its concentration and triggering the Huronian glaciation, possibly the longest episode of glaciation in Earth's history and called snowball Earth.^[11]

Eventually, the evolution of aerobic organisms that consumed oxygen established an equilibrium in its availability. Free oxygen has been an important constituent of the atmosphere ever since.^[11]



O₂ build-up in the Earth's atmosphere. Red and green lines represent the range of the estimates while time is measured in billions of years ago (Ga).

Stage 1 (3.85–2.45 Ga): Practically no O₂ in the atmosphere. The oceans were also largely anoxic with the possible exception of O₂ gases in the shallow oceans.

Stage 2 (2.45–1.85 Ga): O₂ produced, and rose to values of 0.02 and 0.04 atm, but absorbed in oceans and seabed rock.

Stage 3 (1.85–0.85 Ga): O₂ starts to gas out of the oceans, but is absorbed by land surfaces. There was no significant change in terms of oxygen level.

Stages 4 and 5 (0.85–present): O₂ sinks filled and the gas accumulates.^[1]



Cyanobacteria: Responsible for the buildup of oxygen in the earth's atmosphere

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- 3 Late evolution of oxy-photosynthesis theory
- 4 Role in mineral diversification
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Timing

The most widely accepted chronology of the Great Oxygenation Event suggests that free oxygen was first produced by prokaryotic and then later eukaryotic organisms that carried out oxygenic photosynthesis more efficiently. The excess of oxygen those organisms produce is a waste product. These organisms lived long before the GOE,^[12] perhaps as early as 3,500 million years ago.

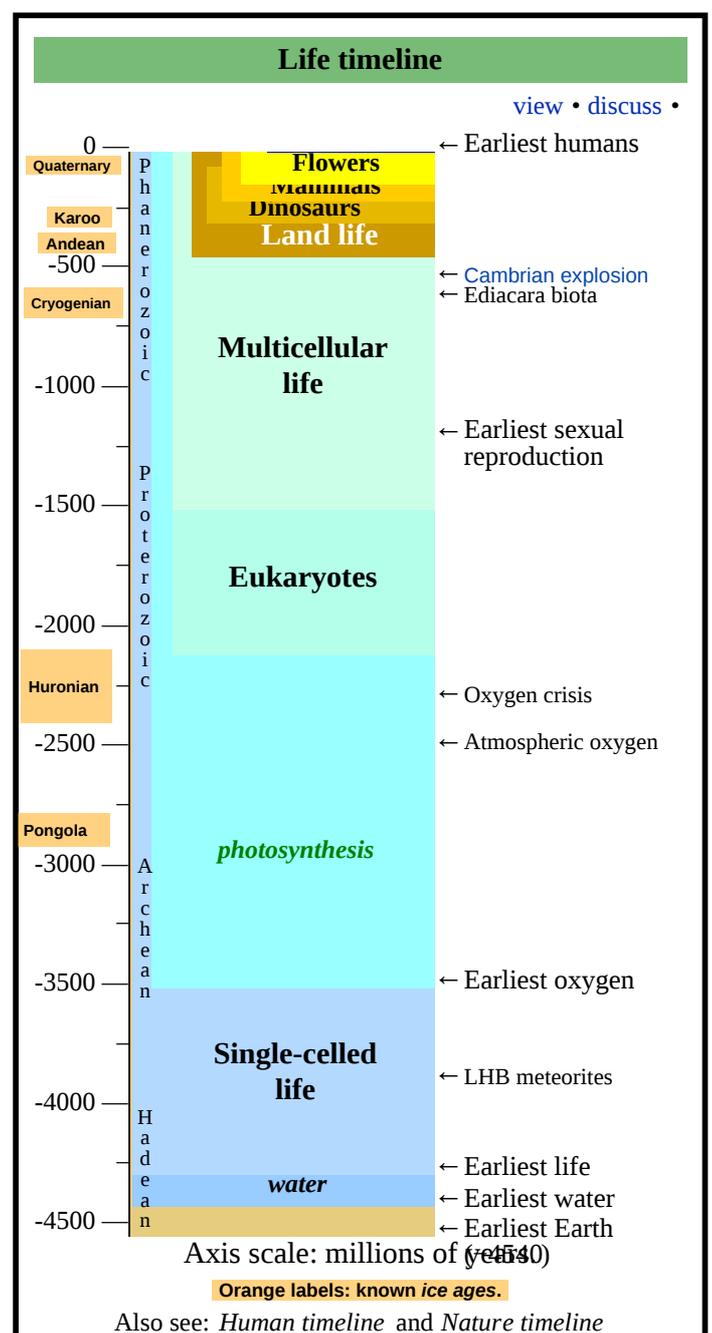
Initially, the oxygen they produced would have quickly been removed from the atmosphere by the chemical weathering of reducing minerals (those able to be oxidized), most notably iron. This 'mass rusting' led to the deposition of iron(III) oxide in the form of banded-iron formations such as the sediments in Minnesota and Pilbara, Western Australia.

Oxygen only began to persist in the atmosphere in small quantities shortly (~50 million years) before the start of the GOE.^[13] Without a mineral sink in the form of iron, oxygen could have accumulated very rapidly.

For example, at today's rates of photosynthesis (which are much greater than those in the land-plant-free Precambrian), modern atmospheric O₂ levels could be produced in around 2,000 years.^[14]

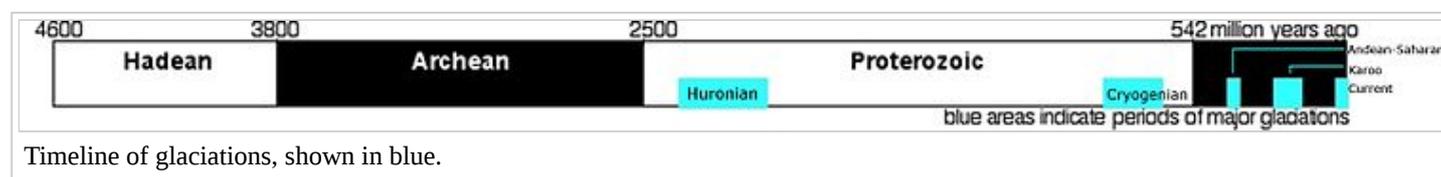
Another hypothesis is that oxygen producers did not evolve until just before the major rise in atmospheric oxygen concentration.^[15] This is based on interpretation of the supposed oxygen indicator, mass-independent fractionation of sulfur isotopes, used in previous studies. This hypothesis would eliminate the need to explain a lag in time between the evolution of oxyphotosynthetic microbes and the rise in free oxygen.

Either way, the oxygen did eventually accumulate in the atmosphere, with two major consequences: First, it oxidized atmospheric methane (a strong greenhouse gas) to carbon dioxide (a weaker one) and water, triggering the Huronian glaciation, 2.4 to 2.1 billion years ago. The latter may have been a full-blown, and possibly the longest ever,



snowball Earth episode, lasting 300–400 million years.^{[15][16]} Second, the increased oxygen concentrations provided a new opportunity for biological diversification, as well as tremendous changes in the nature of chemical interactions between rocks, sand, clay, and other geological substrates and the Earth's air, oceans, and other surface waters.

Despite the natural recycling of organic matter, life had remained energetically limited until the widespread availability of oxygen. This breakthrough in metabolic evolution greatly increased the free energy supply to living organisms, having a truly global environmental impact; mitochondria evolved after the GOE. With more energy available from oxygen, organisms had the means for new, more complex morphologies. These new morphologies in turn helped drive evolution through interaction between organisms.^[17]



Time lag theory

The gap between the start of oxygen production from photosynthetic organisms and the geologically rapid increase in atmospheric oxygen (about 2.5–2.4 billion years ago) may have been as long as 900 million years. Several hypotheses might explain the time lag:

Tectonic trigger

The oxygen increase had to await tectonically driven changes in the Earth, including the appearance of shelf seas, where reduced organic carbon could reach the sediments and be buried.^[18] The newly produced oxygen was first consumed in various chemical reactions in the oceans, primarily with iron. Evidence is found in older rocks that contain massive banded iron formations that were apparently laid down as this iron and oxygen first combined; most of the planet's commercial iron ore is in these deposits. Researchers found that the amount of oxygen in the air spiked each time smaller land masses collided to form a super-continent. These massive pile ups generated mountain chains and as these mountains eroded, they released nutrients into the ocean, feeding the cyanobacteria that carry out photosynthesis.^[19]



2.1 billion year old rock showing banded iron formation

Nickel famine

The early chemosynthetic organisms would have been a source of methane, which is an important trap for molecular oxygen, because oxygen readily oxidizes methane to carbon dioxide (CO_2) and water in the presence of UV radiation. Modern methanogens require nickel as an enzyme cofactor. As the Earth's crust cooled, the supply of nickel from volcanoes was reduced and hence less methane was produced. This allowed the oxygen percentage of the atmosphere to rise as the decrease in production of methane allowed the oxygen producing algae and other such life forms to, in a sense, out-produce the methane producers.^[20] From 2.7 to 2.4 billion years ago, the rate of deposition of nickel declined steadily; it was originally 400 times today's levels.^[21]

Bistability

A theory, called *bistability*, is a mathematical model of the atmosphere. In this model, UV shielding decreases the rate of methane oxidation once oxygen levels are sufficient to support the formation of an ozone layer. This explanation proposes an atmospheric system experiencing two steady states, one with lower (0.02%) atmospheric oxygen content, and the other with higher (21% or more) oxygen content. The Great Oxygenation Event can then be understood as a switch between the lower to the upper stable steady states.^[22]

Hydrogen gas

Another factor is the presence of hydrogen gas. The appearance of cyanobacteria might explain the decline of hydrogen gas and why Earth's air is so oxygen-rich.

Some bacteria living in the early oceans had the ability to separate water into hydrogen and oxygen molecules. By using the energy of the Sun, the separate hydrogen molecules were able to create organic compounds, and oxygen was a by-product. If the hydrogen-heavy compounds were buried, it would have allowed oxygen to accumulate in the atmosphere.

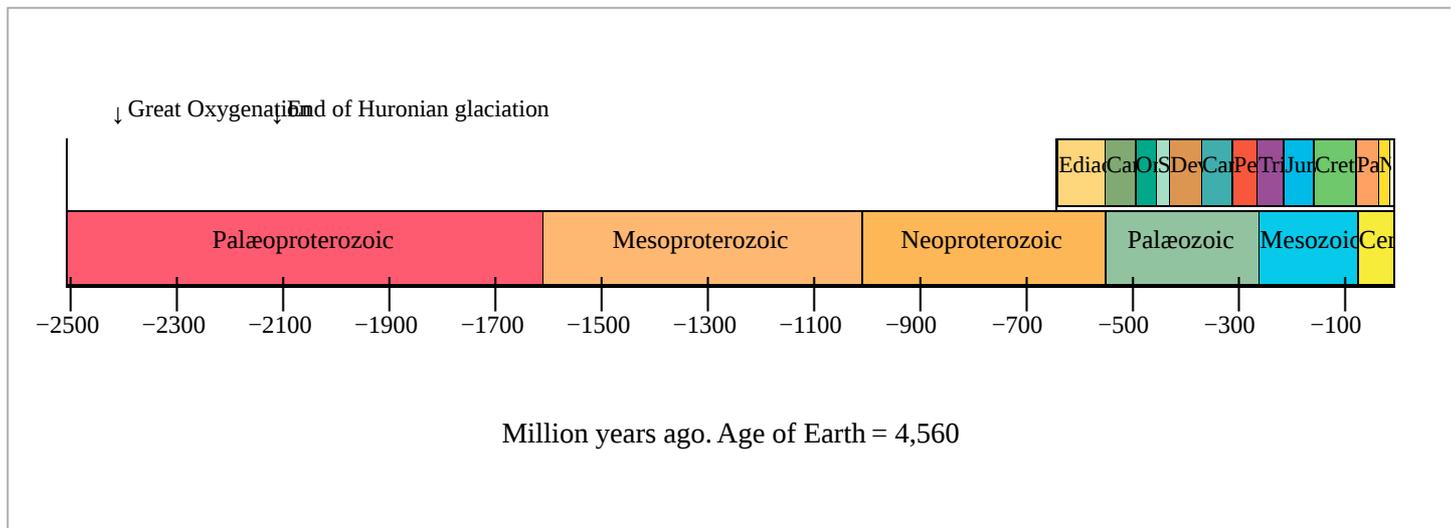
However, in 2001 scientists realized that the hydrogen would instead escape into space through a process called methane photolysis, in which hydrogen-containing methane reacts with oxygen releasing hydrogen. This hypothesis could explain why the early Earth stayed warm enough to sustain oxygen producing lifeforms.^[23]

Late evolution of oxy-photosynthesis theory

There is a possibility that the oxygen indicator was misinterpreted. During the proposed time of the lag in the previous theory, there was a change from mass-independently fractionated (MIF) sulfur to mass-dependently fractionated (MDF) sulfur in sediments. This was assumed to be a result of the appearance of oxygen in the atmosphere (since oxygen would have prevented the photolysis of sulfur dioxide, which causes MIF). This change from MIF to MDF of sulfur isotopes also may have been caused by an increase in glacial weathering, or the homogenization of the marine sulfur pool as a result of an increased thermal gradient during the Huronian glaciation period.^[15]

Role in mineral diversification

The Great Oxygenation Event triggered an explosive growth in the diversity of minerals on Earth. This now meant that many elements could occur in one or more oxidized forms in the near-surface environment.^[24] It is estimated that the Great Oxygenation Event alone was directly responsible for more than 2,500 new minerals of the total of about 4,500 minerals found on Earth. Most of these new minerals were formed after the Great Oxygenation event as hydrated and oxidized forms due to dynamic mantle and crust processes.^[25]



See also

- Geological history of oxygen
- Iodide
- Medea hypothesis
- Pasteur point
- Rare Earth hypothesis

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External links

- First breath: Earth's billion-year struggle for oxygen *New Scientist*, #2746, 5 February 2010 by Nick Lane. [2]

Retrieved from "https://en.wikipedia.org/w/index.php?title=Great_Oxygenation_Event&oldid=787296108"

Categories: Proterozoic | Origin of life | Oxygen | Climate history | Events in the geological history of Earth | Extinction events

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Greenhouse gas

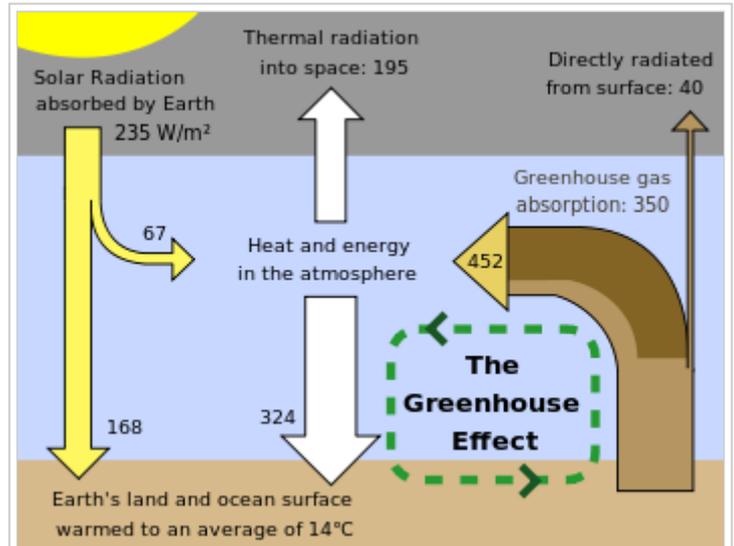
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A **greenhouse gas** (abbrev. **GHG**) is a gas in an atmosphere that absorbs and emits radiation within the thermal infrared range. This process is the fundamental cause of the greenhouse effect.^[1] The primary greenhouse gases in Earth's atmosphere are water vapor, carbon dioxide, methane, nitrous oxide, and ozone. Without greenhouse gases, the average temperature of Earth's surface would be about $-18\text{ }^{\circ}\text{C}$ ($0\text{ }^{\circ}\text{F}$),^[2] rather than the present average of $15\text{ }^{\circ}\text{C}$ ($59\text{ }^{\circ}\text{F}$).^{[3][4][5]} In the Solar System, the atmospheres of Venus, Mars and Titan also contain gases that cause a greenhouse effect.

Human activities since the beginning of the Industrial Revolution (taken as the year 1750) have produced a 40% increase in the atmospheric concentration of carbon dioxide, from 280 ppm in 1750 to 406 ppm in early 2017.^[6] This increase has occurred despite the uptake of a large portion of the

emissions by various natural "sinks" involved in the carbon cycle.^{[7][8]} Anthropogenic carbon dioxide (CO_2) emissions (i.e., emissions produced by human activities) come from combustion of fossil fuels, principally coal, oil, and natural gas, along with deforestation, soil erosion and animal agriculture.^[9]

It has been estimated that if greenhouse gas emissions continue at the present rate, Earth's surface temperature could exceed historical values as early as 2047, with potentially harmful effects on ecosystems, biodiversity and the livelihoods of people worldwide.^[10] Recent estimates suggest that on the current emissions trajectory the Earth could pass a threshold of $2\text{ }^{\circ}\text{C}$ global warming, which the United Nations' IPCC designated as the upper limit to avoid "dangerous" global warming, by 2036.^[11]



Greenhouse effect schematic showing energy flows between space, the atmosphere, and Earth's surface. Energy influx and emittance are expressed in watts per square meter (W/m^2).

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Gases in Earth's atmosphere

Earth's atmosphere is composed mainly of nitrogen and oxygen. The minor constituents, the greenhouse gases and clouds, act to absorb thermal radiation coming from the surface.

Greenhouse gases

Greenhouse gases are those that absorb and emit infrared radiation in the wavelength range emitted by Earth.^[1] In order, the most abundant greenhouse gases in Earth's atmosphere are:

- Water vapor (H₂O)
- Carbon dioxide (CO₂)
- Methane (CH₄)
- Nitrous oxide (N₂O)
- Ozone (O₃)
- Chlorofluorocarbons (CFCs)
- Hydrofluorocarbons (incl. HCFCs and HFCs)

Atmospheric concentrations of greenhouse gases are determined by the balance between sources (emissions of the gas from human activities and natural systems) and sinks (the removal of the gas from the atmosphere by conversion to a different chemical compound).^[12] The proportion of an emission remaining in the atmosphere after a specified time is the "airborne fraction" (AF). The *annual airborne fraction* is the ratio of the atmospheric increase in a given year to that year's total emissions. Over the last 50 years (1956–2006) the annual airborne fraction for CO₂ has been increasing at $0.25 \pm 0.21\%$ /year.^[13]

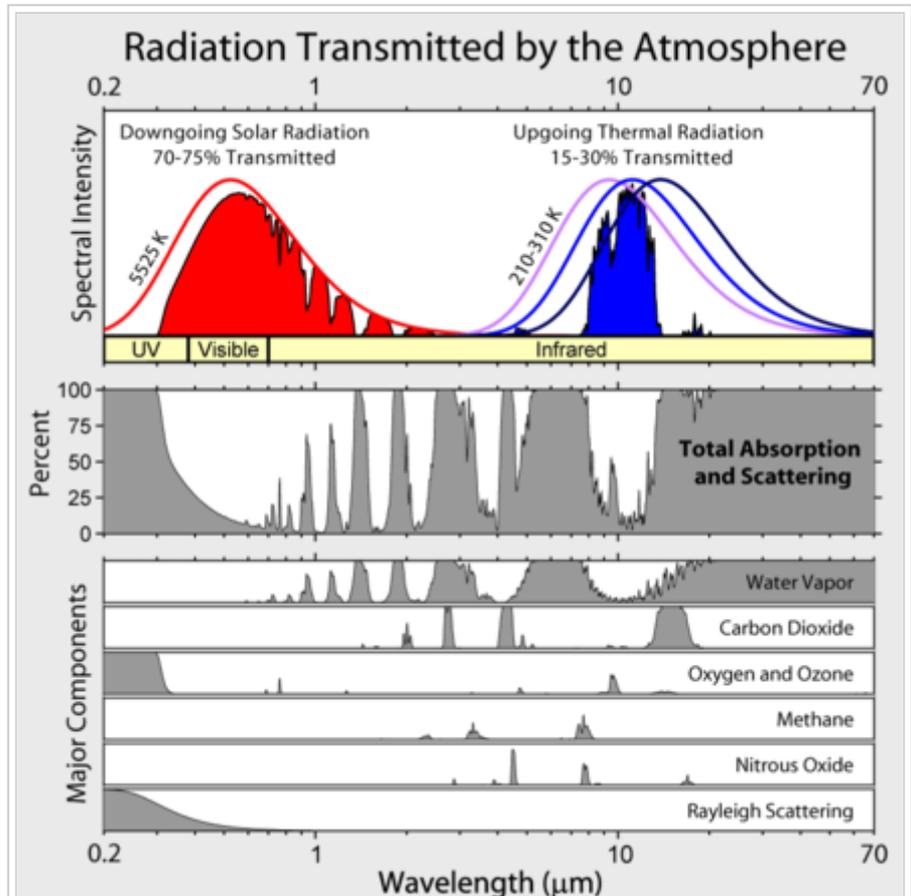
Non-greenhouse gases

The major atmospheric constituents, nitrogen (N₂), oxygen (O₂), and argon (Ar), are not greenhouse gases because molecules containing two atoms of the same element such as N₂ and O₂ and monatomic molecules such as argon (Ar) have no net change in the distribution of their electrical charges when they vibrate. Hence they are almost totally unaffected by infrared radiation. Although molecules containing two atoms of different elements such as carbon monoxide (CO) or hydrogen chloride (HCl) absorb infrared radiation, these molecules are short-lived in the atmosphere owing to their reactivity and solubility. Therefore, they do not contribute significantly to the greenhouse effect and often are omitted when discussing greenhouse gases.

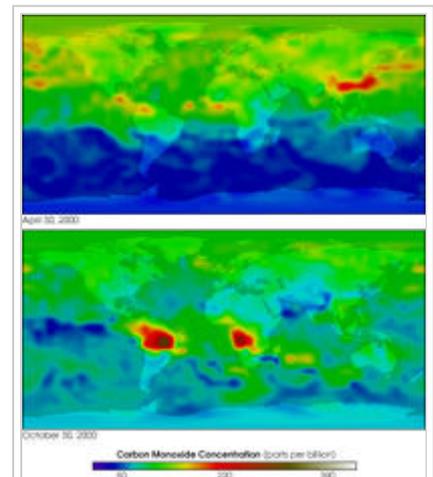
Indirect radiative effects

Some gases have indirect radiative effects (whether or not they are greenhouse gases themselves). This happens in two main ways. One way is that when they break down in the atmosphere they produce another greenhouse gas. For example, methane and carbon monoxide (CO) are oxidized to give carbon dioxide (and methane oxidation also produces water vapor; that will be considered below). Oxidation of CO to CO₂ directly produces an unambiguous increase in radiative forcing although the reason is subtle. The peak of the thermal IR emission from Earth's surface is very close to a strong vibrational absorption band of CO₂ (667 cm⁻¹). On the other hand, the single CO vibrational band only absorbs IR at much higher frequencies (2145 cm⁻¹), where the ~300 K thermal emission of the surface is at least a factor of ten lower. Oxidation of methane to CO₂, which requires reactions with the OH radical, produces an instantaneous reduction, since CO₂ is a weaker greenhouse gas than methane; but it has a longer lifetime. As described below this is not the whole story, since the oxidations of CO and CH₄ are intertwined by both consuming OH radicals. In any case, the calculation of the total radiative effect needs to include both the direct and indirect forcing.

A second type of indirect effect happens when chemical reactions in the atmosphere involving these gases change the concentrations of greenhouse gases. For example, the destruction of non-methane volatile organic compounds (NMVOCs) in the atmosphere can produce ozone. The size of the indirect effect can depend strongly on where and when the gas is emitted.^[15]



Atmospheric absorption and scattering at different wavelengths of electromagnetic waves. The largest absorption band of carbon dioxide is not far from the maximum in the thermal emission from ground, and it partly closes the window of transparency of water; hence its major effect.



The false colors in this image represent concentrations of carbon monoxide in the lower atmosphere, ranging from about 390 parts per billion (dark brown pixels), to 220 parts per billion (red pixels), to 50 parts per billion (blue pixels).^[14]

Methane has a number of indirect effects in addition to forming CO₂. First, the main chemical that destroys methane in the atmosphere is the hydroxyl radical (OH). Methane reacts with OH and so more methane means that the concentration of OH goes down. Effectively, methane increases its own atmospheric lifetime and therefore its overall radiative effect. The second effect is that the oxidation of methane can produce ozone. Third, as well as making CO₂ the oxidation of methane produces water; this is a major source of water vapor in the stratosphere, which is otherwise very dry. CO and NMVOC also produce CO₂ when they are oxidized. They remove OH from the atmosphere and this leads to higher concentrations of methane. The surprising effect of this is that the global warming potential of CO is three times that of CO₂.^[16] The same process that converts NMVOC to carbon dioxide can also lead to the formation of tropospheric ozone. Halocarbons have an indirect effect because they destroy stratospheric ozone. Finally hydrogen can lead to ozone production and CH₄ increases as well as producing water vapor in the stratosphere.^[15]

Contribution of clouds to Earth's greenhouse effect

The major non-gas contributor to Earth's greenhouse effect, clouds, also absorb and emit infrared radiation and thus have an effect on radiative properties of the greenhouse gases. Clouds are water droplets or ice crystals suspended in the atmosphere.^{[17][18]}

Impacts on the overall greenhouse effect

The contribution of each gas to the greenhouse effect is determined by the characteristics of that gas, its abundance, and any indirect effects it may cause. For example, the direct radiative effect of a mass of methane is about 72 times stronger than the same mass of carbon dioxide over a 20-year time frame^[21] but it is present in much smaller concentrations so that its total direct radiative effect is smaller, in part due to its shorter atmospheric lifetime. On the other hand, in addition to its direct radiative impact, methane has a large, indirect radiative effect because it contributes to ozone formation. Shindell *et al.* (2005)^[22] argue that the contribution to climate change from methane is at least double previous estimates as a result of this effect.^[23]

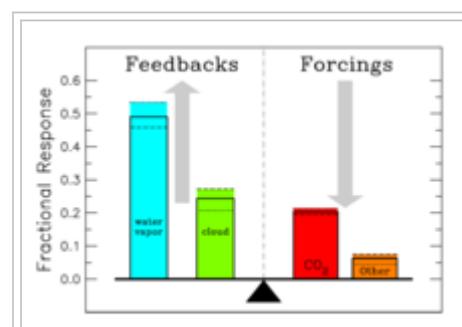
When ranked by their direct contribution to the greenhouse effect, the most important are:^[17]

Compound	Formula	Concentration in atmosphere ^[24] (ppm)	Contribution (%)
Water vapor and clouds	H ₂ O	10–50,000 ^(A)	36–72%
Carbon dioxide	CO ₂	~400	9–26%
Methane	CH ₄	~1.8	4–9%
Ozone	O ₃	2–8 ^(B)	3–7%

notes:

(A) Water vapor strongly varies locally ^[25]

(B) The concentration in stratosphere. About 90% of the ozone in Earth's atmosphere is contained in the stratosphere.



Schmidt *et al.* (2010)^[19] analysed how individual components of the atmosphere contribute to the total greenhouse effect. They estimated that water vapor accounts for about 50% of Earth's greenhouse effect, with clouds contributing 25%, carbon dioxide 20%, and the minor greenhouse gases and aerosols accounting for the remaining 5%. In the study, the reference model atmosphere is for 1980 conditions. Image credit: NASA.^[20]

In addition to the main greenhouse gases listed above, other greenhouse gases include sulfur hexafluoride, hydrofluorocarbons and perfluorocarbons (see IPCC list of greenhouse gases). Some greenhouse gases are not often listed. For example, nitrogen trifluoride has a high global warming potential (GWP) but is only present in very small quantities.^[26]

Proportion of direct effects at a given moment

It is not possible to state that a certain gas causes an exact percentage of the greenhouse effect. This is because some of the gases absorb and emit radiation at the same frequencies as others, so that the total greenhouse effect is not simply the sum of the influence of each gas. The higher ends of the ranges quoted are for each gas alone; the lower ends account for overlaps with the other gases.^{[17][18]} In addition, some gases such as methane are known to have large indirect effects that are still being quantified.^[27]

Atmospheric lifetime

Aside from water vapor, which has a residence time of about nine days,^[28] major greenhouse gases are well mixed and take many years to leave the atmosphere.^[29] Although it is not easy to know with precision how long it takes greenhouse gases to leave the atmosphere, there are estimates for the principal greenhouse gases. Jacob (1999)^[30] defines the lifetime τ of an atmospheric species X in a one-box model as the average time that a molecule of X remains in the box. Mathematically τ can be defined as the ratio of the mass m (in kg) of X in the box to its removal rate, which is the sum of the flow of X out of the box (F_{out}), chemical loss of X (L), and deposition of X (D) (all in kg/s): $\tau = \frac{m}{F_{out} + L + D}$.^[30] If one stopped pouring any of this gas into the box, then after a time τ , its concentration would be about halved.

The atmospheric lifetime of a species therefore measures the time required to restore equilibrium following a sudden increase or decrease in its concentration in the atmosphere. Individual atoms or molecules may be lost or deposited to sinks such as the soil, the oceans and other waters, or vegetation and other biological systems, reducing the excess to background concentrations. The average time taken to achieve this is the mean lifetime.

Carbon dioxide has a variable atmospheric lifetime, and cannot be specified precisely.^[31] The atmospheric lifetime of CO₂ is estimated of the order of 30–95 years.^[32] This figure accounts for CO₂ molecules being removed from the atmosphere by mixing into the ocean, photosynthesis, and other processes. However, this excludes the balancing fluxes of CO₂ into the atmosphere from the geological reservoirs, which have slower characteristic rates.^[33] Although more than half of the CO₂ emitted is removed from the atmosphere within a century, some fraction (about 20%) of emitted CO₂ remains in the atmosphere for many thousands of years.^[34]^[35] ^[36] Similar issues apply to other greenhouse gases, many of which have longer mean lifetimes than CO₂. E.g., N₂O has a mean atmospheric lifetime of 114 years.^[21]

Radiative forcing

Earth absorbs some of the radiant energy received from the sun, reflects some of it as light and reflects or radiates the rest back to space as heat.^[37] Earth's surface temperature depends on this balance between incoming and outgoing energy.^[37] If this energy balance is shifted, Earth's surface becomes warmer or cooler, leading to a variety of changes in global climate.^[37]

A number of natural and man-made mechanisms can affect the global energy balance and force changes in Earth's climate.^[37] Greenhouse gases are one such mechanism.^[37] Greenhouse gases absorb and emit some of the outgoing energy radiated from Earth's surface, causing that heat to be retained in the lower atmosphere.^[37]

As explained above, some greenhouse gases remain in the atmosphere for decades or even centuries, and therefore can affect Earth's energy balance over a long period.^[37] Radiative forcing quantifies the effect of factors that influence Earth's energy balance, including changes in the concentrations of greenhouse gases.^[37] Positive radiative forcing leads to warming by increasing the net incoming energy, whereas negative radiative forcing leads to cooling.^[37]

Global warming potential

The global warming potential (GWP) depends on both the efficiency of the molecule as a greenhouse gas and its atmospheric lifetime. GWP is measured relative to the same **mass** of CO₂ and evaluated for a specific timescale. Thus, if a gas has a high (positive) radiative forcing but also a short lifetime, it will have a large GWP on a 20-year scale but a small one on a 100-year scale. Conversely, if a molecule has a longer atmospheric lifetime than CO₂ its GWP will increase with the timescale considered. Carbon dioxide is defined to have a GWP of 1 over all time periods.

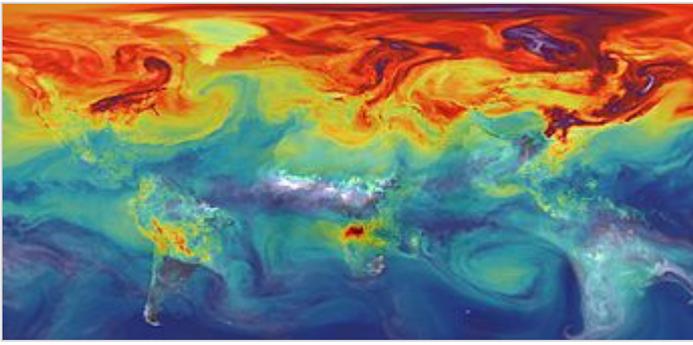
Methane has an atmospheric lifetime of 12 ± 3 years. The 2007 IPCC report lists the GWP as 72 over a time scale of 20 years, 25 over 100 years and 7.6 over 500 years.^[21] A 2014 analysis, however, states that although methane's initial impact is about 100 times greater than that of CO₂, because of the shorter atmospheric lifetime, after six or seven decades, the impact of the two gases is about equal, and from then on methane's relative role continues to decline.^[38] The decrease in GWP at longer times is because methane is degraded to water and CO₂ through chemical reactions in the atmosphere.

Examples of the atmospheric lifetime and GWP relative to CO₂ for several greenhouse gases are given in the following table:^[21]

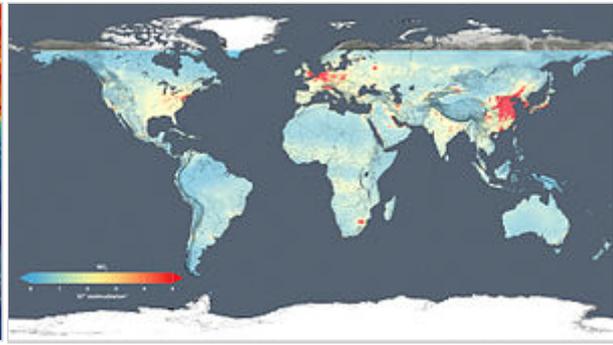
Atmospheric lifetime and GWP relative to CO₂ at different time horizon for various greenhouse gases.

Gas name	Chemical formula	Lifetime (years)	Global warming potential (GWP) for given time horizon		
			20-yr	100-yr	500-yr
Carbon dioxide	CO ₂	30–95	1	1	1
Methane	CH ₄	12	72	25	7.6
Nitrous oxide	N ₂ O	114	289	298	153
CFC-12	CCl ₂ F ₂	100	11 000	10 900	5 200
HCFC-22	CHClF ₂	12	5 160	1 810	549
Tetrafluoromethane	CF ₄	50 000	5 210	7 390	11 200
Hexafluoroethane	C ₂ F ₆	10 000	8 630	12 200	18 200
Sulfur hexafluoride	SF ₆	3 200	16 300	22 800	32 600
Nitrogen trifluoride	NF ₃	740	12 300	17 200	20 700

The use of CFC-12 (except some essential uses) has been phased out due to its ozone depleting properties.^[39] The phasing-out of less active HCFC-compounds will be completed in 2030.^[40]



Carbon dioxide in Earth's atmosphere if *half* of global-warming emissions^{[41][42]} are *not* absorbed. (NASA simulation; 9 November 2015)



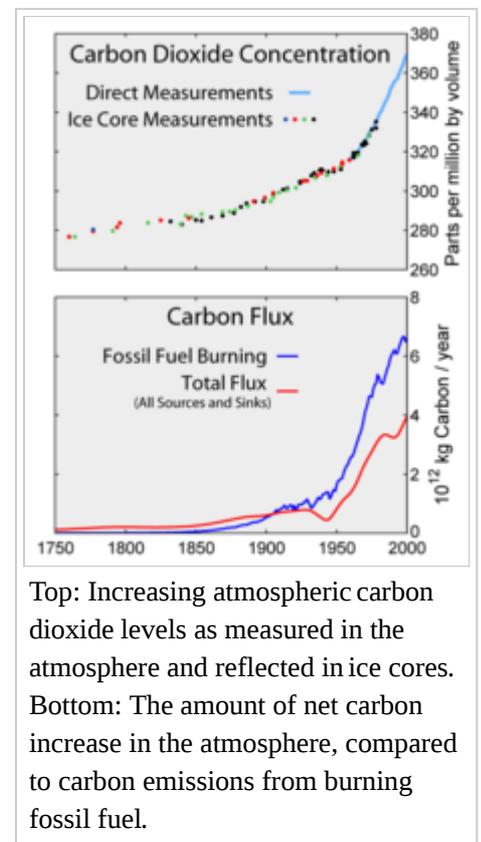
Nitrogen dioxide 2014 – global air quality levels (released 14 December 2015).^[43]

Natural and anthropogenic sources

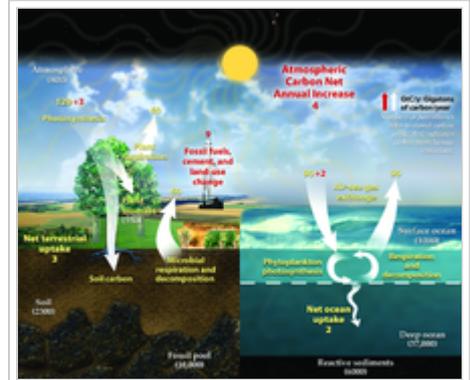
Aside from purely human-produced synthetic halocarbons, most greenhouse gases have both natural and human-caused sources. During the pre-industrial Holocene, concentrations of existing gases were roughly constant. In the industrial era, human activities have added greenhouse gases to the atmosphere, mainly through the burning of fossil fuels and clearing of forests.^{[45][46]}

The 2007 Fourth Assessment Report compiled by the IPCC (AR4) noted that "changes in atmospheric concentrations of greenhouse gases and aerosols, land cover and solar radiation alter the energy balance of the climate system", and concluded that "increases in anthropogenic greenhouse gas concentrations is very likely to have caused most of the increases in global average temperatures since the mid-20th century".^[47] In AR4, "most of" is defined as more than 50%.

Abbreviations used in the two tables below: ppm = parts-per-million; ppb = parts-per-billion; ppt = parts-per-trillion; W/m² = watts per square metre



Top: Increasing atmospheric carbon dioxide levels as measured in the atmosphere and reflected in ice cores. Bottom: The amount of net carbon increase in the atmosphere, compared to carbon emissions from burning fossil fuel.



This diagram shows a simplified representation of the contemporary global carbon cycle. Changes are measured in gigatons of carbon per year (GtC/y). Canadell *et al.* (2007) estimated the growth rate of global average atmospheric CO₂ for 2000–2006 as 1.93 parts-per-million per year (4.1 petagrams of carbon per year).^[44]

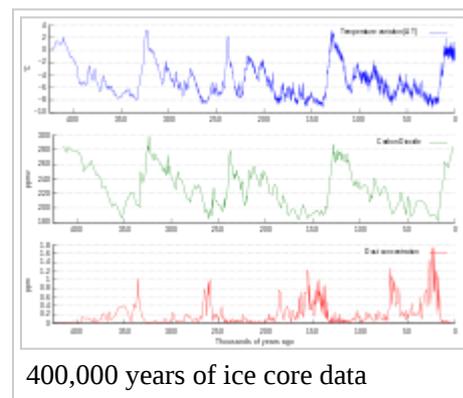
Current greenhouse gas concentrations^[48]

Gas	Pre-1750 tropospheric concentration ^[49]	Recent tropospheric concentration ^[50]	Absolute increase since 1750	Percentage increase since 1750	Increased radiative forcing (W/m ²) ^[51]
Carbon dioxide (CO ₂)	280 ppm ^[52]	395.4 ppm ^[53]	115.4 ppm	41.2%	1.88
Methane (CH ₄)	700 ppb ^[54]	1893 ppb ^[55] 1762 ppb ^[55]	1193 ppb / 1062 ppb	170.4% / 151.7%	0.49
Nitrous oxide (N ₂ O)	270 ppb ^{[51][56]}	326 ppb ^[55] 324 ppb ^[55]	56 ppb / 54 ppb	20.7% / 20.0%	0.17
Tropospheric ozone (O ₃)	237 ppb ^[49]	337 ppb ^[49]	100 ppb	42%	0.4 ^[57]

Relevant to radiative forcing and/or ozone depletion; all of the following have no natural sources and hence zero amounts pre-industrial^[48]

Gas	Recent tropospheric concentration	Increased radiative forcing (W/m ²)
CFC-11 (trichlorofluoromethane) (CCl ₃ F)	236 ppt / 234 ppt	0.061
CFC-12 (CCl ₂ F ₂)	527 ppt / 527 ppt	0.169
CFC-113 (Cl ₂ FC-CClF ₂)	74 ppt / 74 ppt	0.022
HCFC-22 (CHClF ₂)	231 ppt / 210 ppt	0.046
HCFC-141b (CH ₃ CCl ₂ F)	24 ppt / 21 ppt	0.0036
HCFC-142b (CH ₃ CClF ₂)	23 ppt / 21 ppt	0.0042
Halon 1211 (CBrClF ₂)	4.1 ppt / 4.0 ppt	0.0012
Halon 1301 (CBrClF ₃)	3.3 ppt / 3.3 ppt	0.001
HFC-134a (CH ₂ FCF ₃)	75 ppt / 64 ppt	0.0108
Carbon tetrachloride (CCl ₄)	85 ppt / 83 ppt	0.0143
Sulfur hexafluoride (SF ₆)	7.79 ppt ^[58] / 7.39 ppt ^[58]	0.0043
Other halocarbons	Varies by substance	collectively 0.02
Halocarbons in total		0.3574

Ice cores provide evidence for greenhouse gas concentration variations over the past 800,000 years (see the following section). Both CO₂ and CH₄ vary between glacial and interglacial phases, and concentrations of these gases correlate strongly with temperature. Direct data does not exist for periods earlier than those represented in the ice core record, a record that indicates CO₂ mole fractions stayed within a range of 180 ppm to 280 ppm throughout the last 800,000 years, until the increase of the last 250 years. However, various proxies and modeling suggests larger variations in past epochs; 500 million years ago CO₂ levels were likely 10 times higher than now.^[59] Indeed, higher CO₂ concentrations are thought to have prevailed throughout most of the Phanerozoic eon, with concentrations four to six times current concentrations during the Mesozoic era, and ten to fifteen times current concentrations during the early Palaeozoic era until the middle of the Devonian period, about 400 Ma.^{[60][61][62]} The spread of land plants is thought to have reduced CO₂ concentrations during the late Devonian, and plant activities as both sources and sinks of CO₂ have since been important in providing



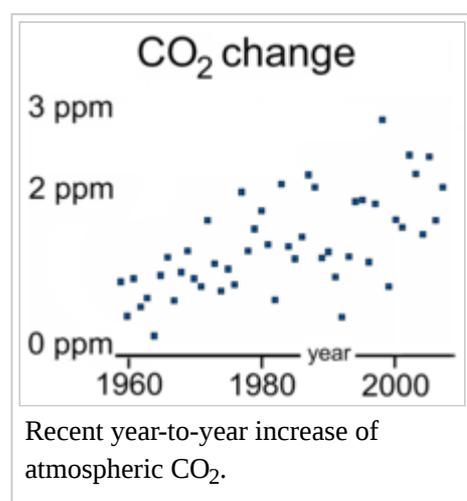
stabilising feedbacks.^[63] Earlier still, a 200-million year period of intermittent, widespread glaciation extending close to the equator (Snowball Earth) appears to have been ended suddenly, about 550 Ma, by a colossal volcanic outgassing that raised the CO₂ concentration of the atmosphere abruptly to 12%, about 350 times modern levels, causing extreme greenhouse conditions and carbonate deposition as limestone at the rate of about 1 mm per day.^[64] This episode marked the close of the Precambrian eon, and was succeeded by the generally warmer conditions of the Phanerozoic, during which multicellular animal and plant life evolved. No volcanic carbon dioxide emission of comparable scale has occurred since. In the modern era, emissions to the atmosphere from volcanoes are only about 1% of emissions from human sources.^{[64][65][66]}

Ice cores

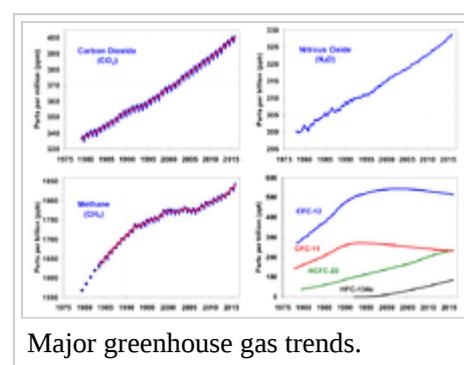
Measurements from Antarctic ice cores show that before industrial emissions started atmospheric CO₂ mole fractions were about 280 parts per million (ppm), and stayed between 260 and 280 during the preceding ten thousand years.^[67] Carbon dioxide mole fractions in the atmosphere have gone up by approximately 35 percent since the 1900s, rising from 280 parts per million by volume to 387 parts per million in 2009. One study using evidence from stomata of fossilized leaves suggests greater variability, with carbon dioxide mole fractions above 300 ppm during the period seven to ten thousand years ago,^[68] though others have argued that these findings more likely reflect calibration or contamination problems rather than actual CO₂ variability.^{[69][70]}

Because of the way air is trapped in ice (pores in the ice close off slowly to form bubbles deep within the firm) and the time period represented in each ice sample analyzed, these figures represent averages of atmospheric concentrations of up to a few centuries rather than annual or decadal levels.

Changes since the Industrial Revolution



Since the beginning of the Industrial Revolution, the concentrations of most of the greenhouse gases have increased. For example, the mole fraction of carbon dioxide has increased from 280 ppm to 400 ppm, or 120 ppm over modern pre-industrial levels. The first 30 ppm increase took place in about 200 years, from the start of the Industrial Revolution to 1958; however the next 90 ppm increase took place within 56 years, from 1958 to 2014.^{[71][72]}



Recent data also shows that the concentration is increasing at a higher rate. In the 1960s, the average annual increase was only 37% of what it was in 2000 through 2007.^[73]

Today, the stock of carbon in the atmosphere increases by more than 3 million tonnes per annum (0.04%) compared with the existing stock. This increase is the result of human activities by burning fossil fuels, deforestation and forest degradation in tropical and boreal regions.^[74]

The other greenhouse gases produced from human activity show similar increases in both amount and rate of increase. Many observations are available online in a variety of Atmospheric Chemistry Observational Databases.

Anthropogenic greenhouse gases

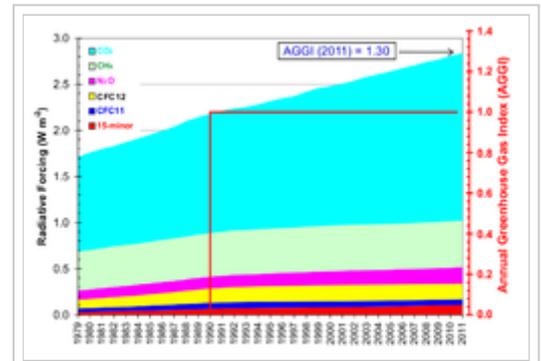
Since about 1750 human activity has increased the concentration of carbon dioxide and other greenhouse gases. Measured atmospheric concentrations of carbon dioxide are currently 100 ppm higher than pre-industrial levels.^[77] Natural sources of carbon dioxide are more than 20 times greater than sources due to human activity,^[78] but over periods longer than a few years natural sources are closely balanced by natural sinks, mainly photosynthesis of carbon compounds by plants and marine plankton. As a result of this balance, the atmospheric mole fraction of carbon dioxide remained between 260 and 280 parts per million for the 10,000 years between the end of the last glacial maximum and the start of the industrial era.^[79]

It is likely that anthropogenic (i.e., human-induced) warming, such as that due to elevated greenhouse gas levels, has had a discernible influence on many physical and biological systems.^[80] Future warming is projected to have a range of impacts, including sea level rise,^[81] increased frequencies and severities of some extreme weather events,^[81] loss of biodiversity,^[82] and regional changes in agricultural productivity.^[82]

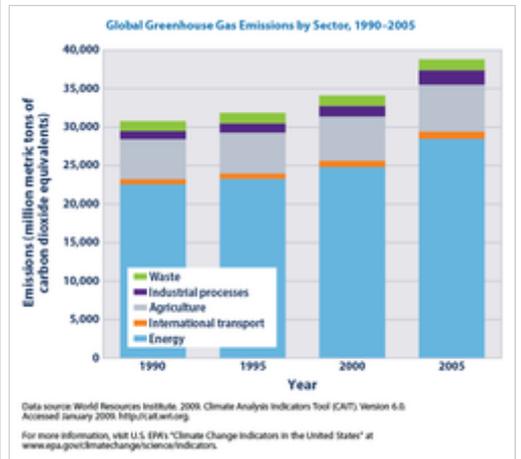
The main sources of greenhouse gases due to human activity are:

- burning of fossil fuels and deforestation leading to higher carbon dioxide concentrations in the air. Land use change (mainly deforestation in the tropics) account for up to one third of total anthropogenic CO₂ emissions.^[79]
- livestock enteric fermentation and manure management,^[83] paddy rice farming, land use and wetland changes, man-made lakes,^[84] pipeline losses, and covered vented landfill emissions leading to higher methane atmospheric concentrations. Many of the newer style fully vented septic systems that enhance and target the fermentation process also are sources of atmospheric methane.
- use of chlorofluorocarbons (CFCs) in refrigeration systems, and use of CFCs and halons in fire suppression systems and manufacturing processes.
- agricultural activities, including the use of fertilizers, that lead to higher nitrous oxide (N₂O) concentrations.

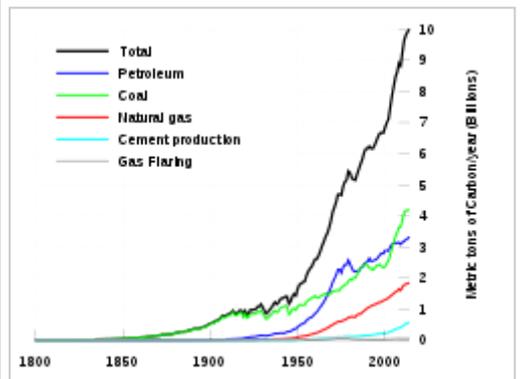
The seven sources of CO₂ from fossil fuel combustion are (with percentage contributions for 2000–2004):^[85]



This graph shows changes in the annual greenhouse gas index (AGGI) between 1979 and 2011.^[75] The AGGI measures the levels of greenhouse gases in the atmosphere based on their ability to cause changes in Earth's climate.^[75]



This bar graph shows global greenhouse gas emissions by sector from 1990 to 2005, measured in carbon dioxide equivalents.^[76]



Modern global CO₂ emissions from the burning of fossil fuels.

Seven main fossil fuel combustion sources	Contribution (%)
Liquid fuels (e.g., gasoline, fuel oil)	36%
Solid fuels (e.g., coal)	35%
Gaseous fuels (e.g., natural gas)	20%
Cement production	3 %
Flaring gas industrially and at wells	< 1%
Non-fuel hydrocarbons	< 1%
"International bunker fuels" of transport not included in national inventories ^[86]	4 %

Carbon dioxide, methane, nitrous oxide (N₂O) and three groups of fluorinated gases (sulfur hexafluoride (SF₆), hydrofluorocarbons (HFCs), and perfluorocarbons (PFCs)) are the major anthropogenic greenhouse gases,^{[87]:147[88]} and are regulated under the Kyoto Protocol international treaty, which came into force in 2005.^[89] Emissions limitations specified in the Kyoto Protocol expired in 2012.^[89] The Cancún agreement, agreed in 2010, includes voluntary pledges made by 76 countries to control emissions.^[90] At the time of the agreement, these 76 countries were collectively responsible for 85% of annual global emissions.^[90]

Although CFCs are greenhouse gases, they are regulated by the Montreal Protocol, which was motivated by CFCs' contribution to ozone depletion rather than by their contribution to global warming. Note that ozone depletion has only a minor role in greenhouse warming though the two processes often are confused in the media. On 15 October 2016, negotiators from over 170 nations meeting at the summit of the United Nations Environment Programme reached a legally binding accord to phase out hydrofluorocarbons (HFCs) in an amendment to the Montreal Protocol.^{[91][92][93]}

Sectors

Tourism

According to UNEP global tourism is closely linked to climate change. Tourism is a significant contributor to the increasing concentrations of greenhouse gases in the atmosphere. Tourism accounts for about 50% of traffic movements. Rapidly expanding air traffic contributes about 2.5% of the production of CO₂. The number of international travelers is expected to increase from 594 million in 1996 to 1.6 billion by 2020, adding greatly to the problem unless steps are taken to reduce emissions.^[94]

Road Haulage

The road haulage industry plays a part in production of CO₂, contributing around 20% of the UK's total carbon emissions a year, with only the energy industry having a larger impact at around 39%.^[95] Average carbon emissions within the haulage industry are falling—in the thirty-year period from 1977 to 2007, the carbon emissions associated with a 200-mile journey fell by 21 percent; NO_x emissions are also down 87 percent, whereas journey times have fallen by around a third.^[96] Due to their size, HGVs often receive criticism regarding their CO₂ emissions; however, rapid development in engine technology and fuel management is having a largely positive effect.

Other

Humans emit, just by breathing, about 450 liter of CO₂ per day (there's an increase of 5% CO₂ in exhaled air compared to inhaled air^{[97][98]}). So, a human generates about 0.9 kg of CO₂/day, or about 238,5 kg of CO₂ in a year. Assuming there are 7,4 billion people on the planet, this means that there is an annual carbon emission of $1,7649 \times 10^9$ tonnes/year

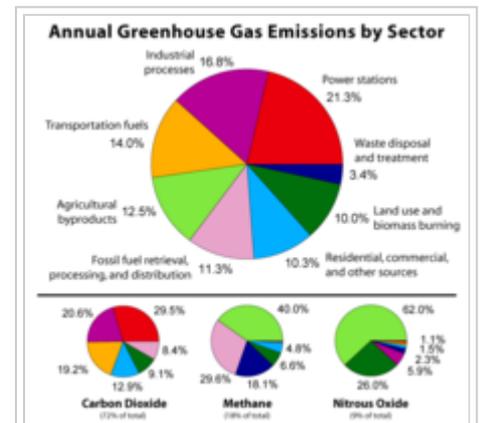
Role of water vapor

Water vapor accounts for the largest percentage of the greenhouse effect, between 36% and 66% for clear sky conditions and between 66% and 85% when including clouds.^[18] Water vapor concentrations fluctuate regionally, but human activity does not directly affect water vapor concentrations except at local scales, such as near irrigated fields. Indirectly, human activity that increases global temperatures will increase water vapor concentrations, a process known as water vapor feedback.^[99] The atmospheric concentration of vapor is highly variable and depends largely on temperature, from less than 0.01% in extremely cold regions up to 3% by mass in saturated air at about 32 °C.^[100] (See Relative humidity#other important facts.)

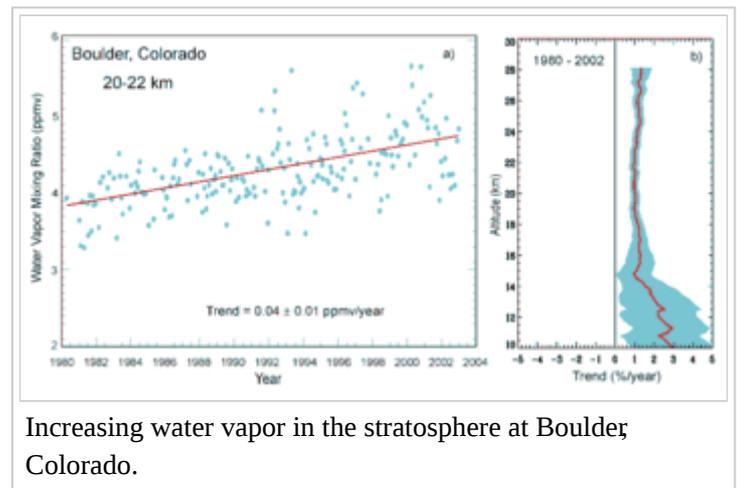
The average residence time of a water molecule in the atmosphere is only about nine days, compared to years or centuries for other greenhouse gases such as CH₄ and CO₂.^[101] Thus, water vapor responds to and amplifies effects of the other greenhouse gases. The Clausius–Clapeyron relation establishes that more water vapor will be present per unit volume at elevated temperatures. This and other basic principles indicate that warming associated with increased concentrations of the other greenhouse gases also will increase the concentration of water vapor (assuming that the relative humidity remains approximately constant; modeling and observational studies find that this is indeed so). Because water vapor is a greenhouse gas, this results in further warming and so is a "positive feedback" that amplifies the original warming. Eventually other earth processes offset these positive feedbacks, stabilizing the global temperature at a new equilibrium and preventing the loss of Earth's water through a Venus-like runaway greenhouse effect.^[102]

Direct greenhouse gas emissions

Between the period 1970 to 2004, GHG emissions (measured in CO₂-equivalent)^[103] increased at an average rate of 1.6% per year, with CO₂ emissions from the use of fossil fuels growing at a rate of 1.9% per year.^{[104][105]} Total anthropogenic emissions at the end of 2009 were estimated at 49.5 gigatonnes CO₂-equivalent.^{[106]:15} These emissions include CO₂ from fossil fuel use and from land use, as well as emissions of methane, nitrous oxide and other GHGs covered by the Kyoto Protocol.



This figure shows the relative fraction of anthropogenic greenhouse gases coming from each of eight categories of sources, as estimated by the Emission Database for Global Atmospheric Research version 3.2, fast track 2000 project [1]. These values are intended to provide a snapshot of global annual greenhouse gas emissions in the year 2000. The top panel shows the sum over all anthropogenic greenhouse gases, weighted by their global warming potential over the next 100 years. This consists of 72% carbon dioxide, 18% methane, 8% nitrous oxide and 1% other gases. Lower panels show the comparable information for each of these three primary greenhouse gases, with the same coloring of sectors as used in the top chart. Segments with less than 1% fraction are not labeled.



Increasing water vapor in the stratosphere at Boulder, Colorado.

At present, the primary source of CO₂ emissions is the burning of coal, natural gas, and petroleum for electricity and heat.^[107]

Regional and national attribution of emissions

There are several different ways of measuring GHG emissions, for example, see World Bank (2010)^{[108]:362} for tables of national emissions data. Some variables that have been reported^[109] include:

- Definition of measurement boundaries: Emissions can be attributed geographically, to the area where they were emitted (the territory principle) or by the activity principle to the territory produced the emissions. These two principles result in different totals when measuring, for example, electricity importation from one country to another, or emissions at an international airport.
- Time horizon of different GHGs: Contribution of a given GHG is reported as a CO₂ equivalent. The calculation to determine this takes into account how long that gas remains in the atmosphere. This is not always known accurately and calculations must be regularly updated to reflect new information.
- What sectors are included in the calculation (e.g., energy industries, industrial processes, agriculture etc.): There is often a conflict between transparency and availability of data.
- The measurement protocol itself: This may be via direct measurement or estimation. The four main methods are the emission factor-based method, mass balance method, predictive emissions monitoring systems, and continuous emissions monitoring systems. These methods differ in accuracy, cost, and usability.

These different measures are sometimes used by different countries to assert various policy/ethical positions on climate change (Banuri *et al.*, 1996, p. 94).^[110] This use of different measures leads to a lack of comparability, which is problematic when monitoring progress towards targets. There are arguments for the adoption of a common measurement tool, or at least the development of communication between different tools.^[109]

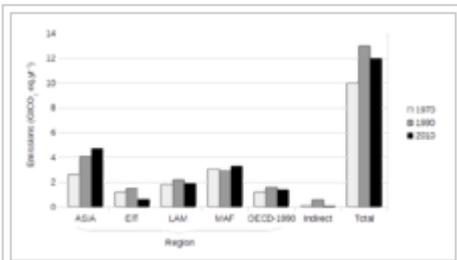
Emissions may be measured over long time periods. This measurement type is called historical or cumulative emissions. Cumulative emissions give some indication of who is responsible for the build-up in the atmospheric concentration of GHGs (IEA, 2007, p. 199).^[111]

The national accounts balance would be positively related to carbon emissions. The national accounts balance shows the difference between exports and imports. For many richer nations, such as the United States, the accounts balance is negative because more goods are imported than they are exported. This is mostly due to the fact that it is cheaper to produce goods outside of developed countries, leading the economies of developed countries to become increasingly dependent on services and not goods. We believed that a positive accounts balance would mean that more production was occurring in a country, so more factories working would increase carbon emission levels.(Holtz-Eakin, 1995, pp.:85;101).^[112]

Emissions may also be measured across shorter time periods. Emissions changes may, for example, be measured against a base year of 1990. 1990 was used in the United Nations Framework Convention on Climate Change (UNFCCC) as the base year for emissions, and is also used in the Kyoto Protocol (some gases are also measured from the year 1995).^{[87]:146,149} A country's emissions may also be reported as a proportion of global emissions for a particular year.

Another measurement is of per capita emissions. This divides a country's total annual emissions by its mid-year population.^{[108]:370} Per capita emissions may be based on historical or annual emissions (Banuri *et al.*, 1996, pp. 106–107).^[110]

Land-use change

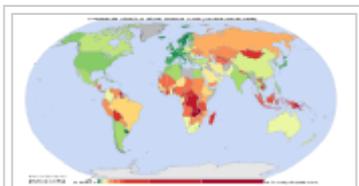


Greenhouse gas emissions from agriculture, forestry and other land use, 1970–2010.

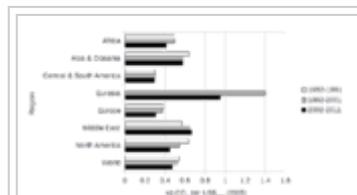
Land-use change, e.g., the clearing of forests for agricultural use, can affect the concentration of GHGs in the atmosphere by altering how much carbon flows out of the atmosphere into carbon sinks.^[113] Accounting for land-use change can be understood as an attempt to measure "net" emissions, i.e., gross emissions from all GHG sources minus the removal of emissions from the atmosphere by carbon sinks (Banuri *et al.*, 1996, pp. 92–93).^[110]

There are substantial uncertainties in the measurement of net carbon emissions.^[114] Additionally, there is controversy over how carbon sinks should be allocated between different regions and over time (Banuri *et al.*, 1996, p. 93).^[110] For instance, concentrating on more recent changes in carbon sinks is likely to favour those regions that have deforested earlier, e.g., Europe.

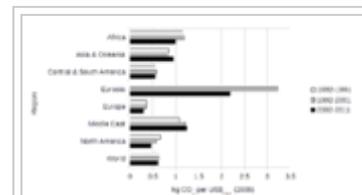
Greenhouse gas intensity



Greenhouse gas intensity in the year 2000, including land-use change.



Carbon intensity of GDP (using PPP) for different regions, 1982–2011.



Carbon intensity of GDP (using MER) for different regions, 1982–2011.

Greenhouse gas intensity is a ratio between greenhouse gas emissions and another metric, e.g., gross domestic product (GDP) or energy use. The terms "carbon intensity" and "emissions intensity" are also sometimes used.^[115] GHG intensities may be calculated using market exchange rates (MER) or purchasing power parity (PPP) (Banuri *et al.*, 1996, p. 96).^[110] Calculations based on MER show large differences in intensities between developed and developing countries, whereas calculations based on PPP show smaller differences.

Cumulative and historical emissions

Cumulative anthropogenic (i.e., human-emitted) emissions of CO₂ from fossil fuel use are a major cause of global warming,^[116] and give some indication of which countries have contributed most to human-induced climate change.^{[117]:15}

Top-5 historic CO₂ contributors by region over the years 1800 to 1988 (in %)

Region	Industrial CO ₂	Total CO ₂
OECD North America	33.2	29.7
OECD Europe	26.1	16.6
Former USSR	14.1	12.5
China	5.5	6.0
Eastern Europe	5.5	4.8

The table above to the left is based on Banuri *et al.* (1996, p. 94).^[110] Overall, developed countries accounted for 83.8% of industrial CO₂ emissions over this time period, and 67.8% of total CO₂ emissions. Developing countries accounted for industrial CO₂ emissions of 16.2% over this time period, and 32.2% of total CO₂ emissions. The estimate of total CO₂ emissions includes biotic carbon emissions, mainly from deforestation. Banuri *et al.* (1996, p. 94)^[110] calculated per capita cumulative emissions based on then-current population. The ratio in per capita emissions between industrialized countries and developing countries was estimated at more than 10 to 1.

Including biotic emissions brings about the same controversy mentioned earlier regarding carbon sinks and land-use change (Banuri *et al.*, 1996, pp. 93–94).^[110] The actual calculation of net emissions is very complex, and is affected by how carbon sinks are allocated between regions and the dynamics of the climate system.

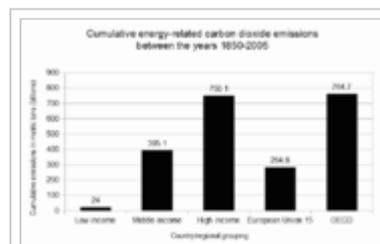
Non-OECD countries accounted for 42% of cumulative energy-related CO₂ emissions between 1890 and 2007.^{[118]:179–180} Over this time period, the US accounted for 28% of emissions; the EU, 23%; Russia, 11%; China, 9%; other OECD countries, 5%; Japan, 4%; India, 3%; and the rest of the world, 18%.^{[118]:179–180}

Changes since a particular base year

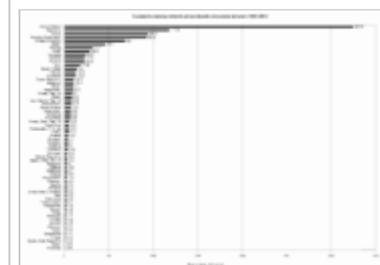
Between 1970 and 2004, global growth in annual CO₂ emissions was driven by North America, Asia, and the Middle East.^[119] The sharp acceleration in CO₂ emissions since 2000 to more than a 3% increase per year (more than 2 ppm per year) from 1.1% per year during the 1990s is attributable to the lapse of formerly declining trends in carbon intensity of both developing and developed nations. China was responsible for most of global growth in emissions during this period. Localised plummeting emissions associated with the collapse of the Soviet Union have been followed by slow emissions growth in this region due to more efficient energy use, made necessary by the increasing proportion of it that is exported.^[85] In comparison, methane has not increased appreciably, and N₂O by 0.25% y⁻¹.

Using different base years for measuring emissions has an effect on estimates of national contributions to global warming.^{[117]:17–18[120]} This can be calculated by dividing a country's highest contribution to global warming starting from a particular base year, by that country's minimum contribution to global warming starting from a particular base year. Choosing between different base years of 1750, 1900, 1950, and 1990 has a significant effect for most countries.^{[117]:17–18} Within the G8 group of countries, it is most significant for the UK, France and Germany. These countries have a long history of CO₂ emissions (see the section on Cumulative and historical emissions).

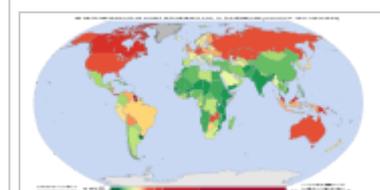
Annual emissions



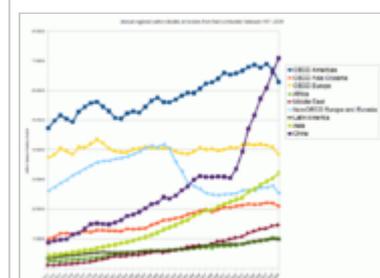
Cumulative energy-related CO₂ emissions between the years 1850–2005 grouped into low-income, middle-income, high-income, the EU-15, and the OECD countries.



Cumulative energy-related CO₂ emissions between the years 1850–2005 for individual countries.

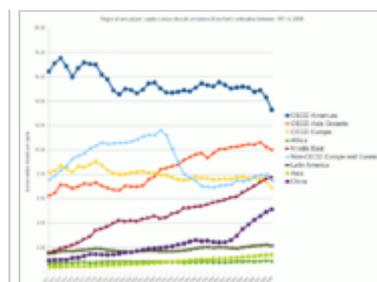


Map of cumulative per capita anthropogenic atmospheric CO₂ emissions by country. Cumulative emissions include land use change, and are measured between the years 1950 and 2000.



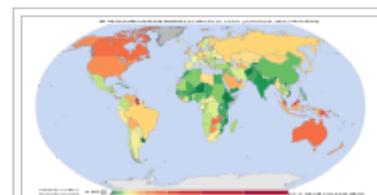
Regional trends in annual CO₂ emissions from fuel combustion between 1971 and 2009.

Annual per capita emissions in the industrialized countries are typically as much as ten times the average in developing countries.^{[87]:144} Due to China's fast economic development, its annual per capita emissions are quickly approaching the levels of those in the Annex I group of the Kyoto Protocol (i.e., the developed countries excluding the USA).^[121] Other countries with fast growing emissions are South Korea, Iran, and Australia (which apart from the oil rich Persian Gulf states, now has the highest per capita emission rate in the world). On the other hand, annual per capita emissions of the EU-15 and the USA are gradually decreasing over time.^[121] Emissions in Russia and Ukraine have decreased fastest since 1990 due to economic restructuring in these countries.^[122]



Regional trends in annual per capita CO₂ emissions from fuel combustion between 1971 and 2009.

Energy statistics for fast growing economies are less accurate than those for the industrialized countries. For China's annual emissions in 2008, the Netherlands Environmental Assessment Agency estimated an uncertainty range of about 10%.^[121]



Per capita anthropogenic greenhouse gas emissions by country for the year 2000 including land-use change.

The GHG footprint, or greenhouse gas footprint, refers to the amount of GHG that are emitted during the creation of products or services. It is more comprehensive than the commonly used carbon footprint, which measures only carbon dioxide, one of many greenhouse gases.

2015 was the first year to see both total global economic growth and a reduction of carbon emissions.^[123]

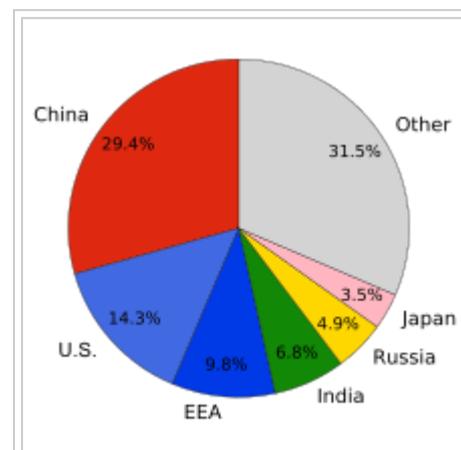
Top emitter countries

Annual

In 2009, the annual top ten emitting countries accounted for about two-thirds of the world's annual energy-related CO₂ emissions.^[124]

Top-10 annual energy-related CO₂ emitters for the year 2009

Country	% of global total annual emissions	Tonnes of GHG per capita
People's Rep. of China	23.6	5.1
United States	17.9	16.9
India	5.5	1.4
Russian Federation	5.3	10.8
Japan	3.8	8.6
Germany	2.6	9.2
Islamic Rep. of Iran	1.8	7.3
Canada	1.8	15.4
South Korea	1.8	10.6
United Kingdom	1.6	7.5



Global carbon dioxide emissions by country.

Cumulative

Top-10 cumulative energy-related CO₂ emitters between 1850 and 2008

Country	% of world total	Metric tonnes CO ₂ per person
United States	28.5	1,132.7
China	9.36	85.4
Russian Federation	7.95	677.2
Germany	6.78	998.9
United Kingdom	5.73	1,127.8
Japan	3.88	367
France	2.73	514.9
India	2.52	26.7
Canada	2.17	789.2
Ukraine	2.13	556.4

Embedded emissions

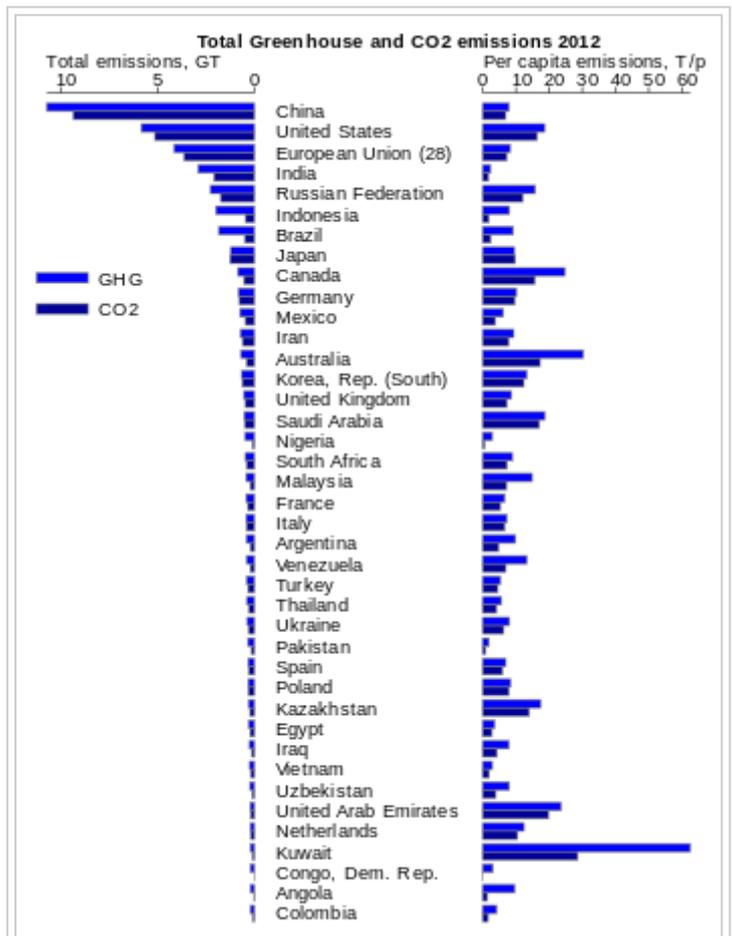
One way of attributing greenhouse gas (GHG) emissions is to measure the embedded emissions (also referred to as "embodied emissions") of goods that are being consumed. Emissions are usually measured according to production, rather than consumption.^[125] For example, in the main international treaty on climate change (the UNFCCC), countries report on emissions produced within their borders, e.g., the emissions produced from burning fossil fuels.^{[118]:179[126]:1} Under a production-based accounting of emissions, embedded emissions on imported goods are attributed to the exporting, rather than the importing, country. Under a consumption-based accounting of emissions, embedded emissions on imported goods are attributed to the importing country, rather than the exporting, country.

Davis and Caldeira (2010)^{[126]:4} found that a substantial proportion of CO₂ emissions are traded internationally. The net effect of trade was to export emissions from China and other emerging markets to consumers in the US, Japan, and Western Europe. Based on annual emissions data from the year 2004, and on a per-capita consumption basis, the top-5 emitting countries were found to be (in tCO₂ per person, per year):

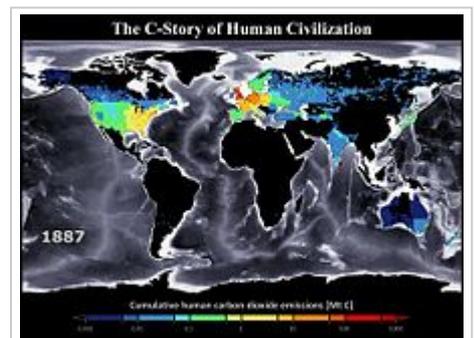
Luxembourg (34.7), the US (22.0), Singapore (20.2), Australia (16.7), and Canada (16.6).^{[126]:5} Carbon Trust research revealed that

approximately 25% of all CO₂ emissions from human activities 'flow' (i.e., are imported or exported) from one country to another. Major developed economies were found to be typically net importers of embodied carbon emissions — with UK consumption emissions 34% higher than production emissions, and Germany (29%), Japan (19%) and the USA (13%) also significant net importers of embodied emissions.^[127]

Effect of policy



The top 40 countries emitting all greenhouse gases, showing both that derived from all sources including land clearance and forestry and also the CO₂ component excluding those sources. Per capita figures are included. Data taken from World Resources Institute, Washington. Note that Indonesia and Brazil show very much higher than on graphs simply showing fossil fuel use.



Play media

The C-Story of Human Civilization by PIK

Governments have taken action to reduce GHG emissions (climate change mitigation). Assessments of policy effectiveness have included work by the Intergovernmental Panel on Climate Change,^[128] International Energy Agency,^{[129][130]} and United Nations Environment Programme.^[131] Policies implemented by governments have included^{[132][133][134]} national and regional targets to reduce emissions, promoting energy efficiency, and support for renewable energy such as Solar energy as an effective use of renewable energy because solar uses energy from the sun and does not release pollutants into the air.

Countries and regions listed in Annex I of the United Nations Framework Convention on Climate Change (UNFCCC) (i.e., the OECD and former planned economies of the Soviet Union) are required to submit periodic assessments to the UNFCCC of actions they are taking to address climate change.^{[134]:3} Analysis by the UNFCCC (2011)^{[134]:8} suggested that policies and measures undertaken by Annex I Parties may have produced emission savings of 1.5 thousand Tg CO₂-eq in the year 2010, with most savings made in the energy sector. The projected emissions saving of 1.5 thousand Tg CO₂-eq is measured against a hypothetical "baseline" of Annex I emissions, i.e., projected Annex I emissions in the absence of policies and measures. The total projected Annex I saving of 1.5 thousand CO₂-eq does not include emissions savings in seven of the Annex I Parties.^{[134]:8}

Projections

A wide range of projections of future GHG emissions have been produced.^[135] Rogner *et al.* (2007)^[136] assessed the scientific literature on GHG projections. Rogner *et al.* (2007)^[104] concluded that unless energy policies changed substantially, the world would continue to depend on fossil fuels until 2025–2030. Projections suggest that more than 80% of the world's energy will come from fossil fuels. This conclusion was based on "much evidence" and "high agreement" in the literature.^[104] Projected annual energy-related CO₂ emissions in 2030 were 40–110% higher than in 2000, with two-thirds of the increase originating in developing countries.^[104] Projected annual per capita emissions in developed country regions remained substantially lower (2.8–5.1 tonnes CO₂) than those in developed country regions (9.6–15.1 tonnes CO₂).^[137] Projections consistently showed increase in annual world GHG emissions (the "Kyoto" gases,^[138] measured in CO₂-equivalent) of 25–90% by 2030, compared to 2000.^[104]

Relative CO₂ emission from various fuels

One liter of gasoline, when used as a fuel, produces 2.32 kg (about 1300 liters or 1.3 cubic meters) of carbon dioxide, a greenhouse gas. One US gallon produces 19.4 lb (1,291.5 gallons or 172.65 cubic feet)^{[139][140][141]}

**Mass of carbon dioxide emitted per quantity of energy
for various fuels^[142]**

Fuel name	CO ₂ emitted (lbs/10 ⁶ Btu)	CO ₂ emitted (g/MJ)	CO ₂ emitted (g/KWh)
Natural gas	117	50.30	181.08
Liquefied petroleum gas	139	59.76	215.14
Propane	139	59.76	215.14
Aviation gasoline	153	65.78	236.81
Automobile gasoline	156	67.07	241.45
Kerosene	159	68.36	246.10
Fuel oil	161	69.22	249.19
Tires/tire derived fuel	189	81.26	292.54
Wood and wood waste	195	83.83	301.79
Coal (bituminous)	205	88.13	317.27
Coal (sub-bituminous)	213	91.57	329.65
Coal (lignite)	215	92.43	332.75
Petroleum coke	225	96.73	348.23
Tar-sand Bitumen			
Coal (anthracite)	227	97.59	351.32

Life-cycle greenhouse-gas emissions of energy sources

A literature review of numerous energy sources CO₂ emissions by the IPCC in 2011, found that, the CO₂ emission value that fell within the 50th percentile of all total life cycle emissions studies conducted was as follows.^[143]

Lifecycle greenhouse gas emissions by electricity source.

Technology	Description	50th percentile (g CO ₂ /kWh _e)
Hydroelectric	reservoir	4
Ocean Energy	wave and tidal	8
Wind	onshore	12
Nuclear	various generation II reactor types	16
Biomass	various	18
Solar thermal	parabolic trough	22
Geothermal	hot dry rock	45
Solar PV	Polycrystalline silicon	46
Natural gas	various combined cycle turbines without scrubbing	469
Coal	various generator types without scrubbing	1001

Removal from the atmosphere ("sinks")

Natural processes

Greenhouse gases can be removed from the atmosphere by various processes, as a consequence of:

- a physical change (condensation and precipitation remove water vapor from the atmosphere).
- a chemical reaction within the atmosphere. For example, methane is oxidized by reaction with naturally occurring hydroxyl radical, OH• and degraded to CO₂ and water vapor (CO₂ from the oxidation of methane is not included in the methane Global warming potential). Other chemical reactions include solution and solid phase chemistry occurring in atmospheric aerosols.
- a physical exchange between the atmosphere and the other compartments of the planet. An example is the mixing of atmospheric gases into the oceans.
- a chemical change at the interface between the atmosphere and the other compartments of the planet. This is the case for CO₂, which is reduced by photosynthesis of plants, and which, after dissolving in the oceans, reacts to form carbonic acid and bicarbonate and carbonate ions (see ocean acidification).
- a photochemical change. Halocarbons are dissociated by UV light releasing Cl• and F• as free radicals in the stratosphere with harmful effects on ozone (halocarbons are generally too stable to disappear by chemical reaction in the atmosphere).

Negative emissions

A number of technologies remove greenhouse gases emissions from the atmosphere. Most widely analysed are those that remove carbon dioxide from the atmosphere, either to geologic formations such as bio-energy with carbon capture and storage^{[144][145][146]} and carbon dioxide air capture,^[146] or to the soil as in the case with biochar.^[146] The IPCC has pointed out that many long-term climate scenario models require large-scale manmade negative emissions to avoid serious climate change.^[147]

History of scientific research

In the late 19th century scientists experimentally discovered that N₂ and O₂ do not absorb infrared radiation (called, at that time, "dark radiation"), while water (both as true vapor and condensed in the form of microscopic droplets suspended in clouds) and CO₂ and other poly-atomic gaseous molecules do absorb infrared radiation. In the early 20th century researchers realized that greenhouse gases in the atmosphere made Earth's overall temperature higher than it would be without them. During the late 20th century, a scientific consensus evolved that increasing concentrations of greenhouse gases in the atmosphere cause a substantial rise in global temperatures and changes to other parts of the climate system,^[148] with consequences for the environment and for human health.

See also

- Attribution of recent climate change
- Carbon accounting
- Carbon credit
- Carbon emissions reporting
- Carbon neutrality
- Carbon offset
- Cap and Trade
- Deforestation and climate change
- Effects of global warming
- Emission standard
- Environmental impact of aviation
- Greenhouse debt
- Hydrogen economy
- Integrated Carbon Observation System
- List of countries by electricity production from renewable sources
- List of international environmental agreements
- Low-carbon economy
- Methane
- Mobile source air pollution
- Perfluorotributylamine
- Physical properties of greenhouse gases
- Sustainability measurement
- World energy consumption
- Zero-emissions vehicle

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50. Because atmospheric concentrations of most gases tend to vary systematically over the course of a year, figures given represent averages over a 12-month period for all gases except ozone (O₃), for which a current global value has been estimated (IPCC, 2001, *Table 4.1a*). CO₂ averages for year 2012 are taken from the National Oceanic and Atmospheric Administration, Earth System Research Laboratory web site: www.esrl.noaa.gov/gmd/ccgg/trends maintained by Dr Pieter Tans. For other chemical species, the values given are averages for 2011. These data are found on the CDIAC AGAGE web site: <http://cdiac.ornl.gov/ndps/alegage.htm> or the AGAGE home page: <http://agage.eas.gatech.edu>
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53. Recent CO₂ concentration (395.4 ppm) is the 2013 average taken from globally averaged marine surface data given by the National Oceanic and Atmospheric Administration Earth System Research Laboratory website: <http://www.esrl.noaa.gov/gmd/ccgg/trends/index.html#global> Please read the material on that web page and reference Dr. Pieter Tans when citing this average (Dr Pieter Tans, NOAA/ESRL <http://www.esrl.noaa.gov/gmd/ccgg/trends/>). The oft-cited Mauna Loa average for 2012 is 393.8 ppm, which is a good approximation although typically about 1 ppm higher than the spatial average given above. Refer to <http://www.esrl.noaa.gov/gmd/ccgg/trends/> for records back to the late 1950s.
54. ppb = parts-per-billion
55. The first value in a cell represents Mace Head, Ireland, a mid-latitude Northern-Hemisphere site, and the second value represents Cape Grim, Tasmania, a mid-latitude Southern-Hemisphere site. "Current" values given for these gases are annual arithmetic averages based on monthly background concentrations for year 2011. The SF₆ values are from the AGAGE gas chromatography – mass spectrometer (gc-ms) Medusa measuring system. Source: Advanced Global Atmospheric Gases Experiment (AGAGE) data posted on CDIAC web site at: http://cdiac.ornl.gov/ftp/ale_gage_Agage/ These data are compiled from data on finer time scales in the ALE/GAGE/AGAGE database [2] (<http://cdiac.ornl.gov/ndps/alegage.html>) (Prinn et al., 2000). These data represent the work of several investigators at various institutions; guidelines on citing the various parts of the AGAGE database are found within the ALE/GAGE/AGAGE database, see [3] (http://cdiac.ornl.gov/ftp/ale_gage_Agage/)
56. The pre-1750 value for N₂O is consistent with ice-core records from 10,000 B.C.E. through 1750 C.E. "Summary for policymakers", *Figure SPM.1* (http://www.ipcc.ch/publications_and_data/ar4/wg1/en/spmssp-human-and.html) IPCC, in IPCC AR4 WG1 2007, p. 3. Referred to by: Blasing, T. J. (February 2013), *Current Greenhouse Gas Concentrations* (http://cdiac.ornl.gov/pns/current_ghg.html) doi:10.3334/CDIAC/atg.032 (<https://doi.org/10.3334/2FCDIAC%2Fatg.032>) on CDIAC 2013
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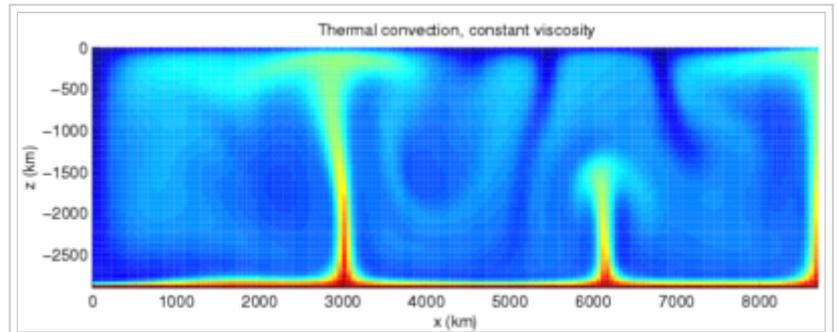
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Heat transfer

From Wikipedia, the free encyclopedia

Heat transfer is a discipline of thermal engineering that concerns the generation, use, conversion, and exchange of thermal energy and heat between physical systems. Heat transfer is classified into various mechanisms, such as thermal conduction, thermal convection, thermal radiation, and transfer of energy by phase changes. Engineers also consider the transfer of mass of differing chemical species, either cold or hot, to achieve heat transfer. While these mechanisms have distinct characteristics, they often occur simultaneously in the same system.



Simulation of thermal convection in the Earth's mantle. Colors span from red and green to blue with decreasing temperatures. A hot, less-dense lower boundary layer sends plumes of hot material upwards, and cold material from the top moves downwards.

Heat conduction, also called diffusion, is the direct microscopic exchange of kinetic energy of particles through the boundary between two systems. When an object is at a different temperature from another body or its surroundings, heat flows so that the body and the surroundings reach the same temperature, at which point they are in thermal equilibrium. Such spontaneous heat transfer always occurs from a region of high temperature to another region of lower temperature, as described by the second law of thermodynamics.

Heat convection occurs when bulk flow of a fluid (gas or liquid) carries heat along with the flow of matter in the fluid. The flow of fluid may be forced by external processes, or sometimes (in gravitational fields) by buoyancy forces caused when thermal energy expands the fluid (for example in a fire plume), thus influencing its own transfer. The latter process is often called "natural convection". All convective processes also move heat partly by diffusion, as well. Another form of convection is forced convection. In this case the fluid is forced to flow by use of a pump, fan or other mechanical means.

Thermal radiation occurs through a vacuum or any transparent medium (solid or fluid). It is the transfer of energy by means of photons in electromagnetic waves governed by the same laws.^[1]

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Overview

Heat is defined in physics as the transfer of thermal energy across a well-defined boundary around a thermodynamic system. The thermodynamic free energy is the amount of work that a thermodynamic system can perform. Enthalpy is a thermodynamic potential, designated by the letter "H", that is the sum of the internal energy of the system (U) plus the product of pressure (P) and volume (V). Joule is a unit to quantify energy, work, or the amount of heat.

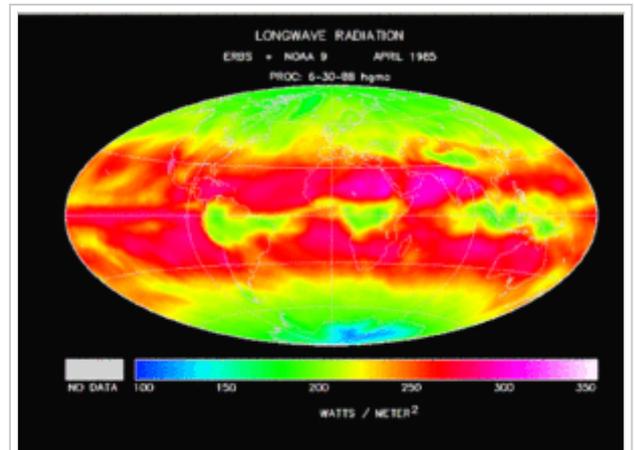
Heat transfer is a process function (or path function), as opposed to functions of state; therefore, the amount of heat transferred in a thermodynamic process that changes the state of a system depends on how that process occurs, not only the net difference between the initial and final states of the process.

Thermodynamic and mechanical heat transfer is calculated with the heat transfer coefficient, the proportionality between the heat flux and the thermodynamic driving force for the flow of heat. Heat flux is a quantitative, vectorial representation of heat-flow through a surface.^[2]

In engineering contexts, the term *heat* is taken as synonymous to thermal energy. This usage has its origin in the historical interpretation of heat as a fluid (*Caloric*) that can be transferred by various causes,^[3] and that is also common in the language of laymen and everyday life.

The transport equations for thermal energy (Fourier's law), mechanical momentum (Newton's law for fluids), and mass transfer (Fick's laws of diffusion) are similar,^{[4][5]} and analogies among these three transport processes have been developed to facilitate prediction of conversion from any one to the others.^[5]

Thermal engineering concerns the generation, use, conversion, and exchange of heat transfer. As such, heat transfer is involved in almost every sector of the economy.^[6] Heat transfer is classified into various mechanisms, such as thermal conduction, thermal convection, thermal radiation, and transfer of energy by phase changes.



Earth's long wave thermal radiation intensity, from clouds, atmosphere and surface.

Mechanisms

The fundamental modes of heat transfer are:

Advection

Advection is the transport mechanism of a fluid from one location to another, and is dependent on motion and momentum of that fluid.

Conduction or diffusion

The transfer of energy between objects that are in physical contact. Thermal conductivity is the property of a material to conduct heat and evaluated primarily in terms of Fourier's Law for heat conduction.

Convection

The transfer of energy between an object and its environment, due to fluid motion. The average temperature is a reference for evaluating properties related to convective heat transfer.

Radiation

The transfer of energy by the emission of electromagnetic radiation.

Advection

By transferring matter, energy—including thermal energy—is moved by the physical transfer of a hot or cold object from one place to another.^[7] This can be as simple as placing hot water in a bottle and heating a bed, or the movement of an iceberg in changing ocean currents. A practical example is thermal hydraulics. This can be described by the formula:

$$Q = v\rho c_p \Delta T$$

where Q is heat flux (W/m^2), ρ is density (kg/m^3), c_p is heat capacity at constant pressure ($\text{J}/\text{kg}\cdot\text{K}$), ΔT is the change in temperature (K), v is velocity (m/s).

Conduction

On a microscopic scale, heat conduction occurs as hot, rapidly moving or vibrating atoms and molecules interact with neighboring atoms and molecules, transferring some of their energy (heat) to these neighboring particles. In other words, heat is transferred by conduction when adjacent atoms vibrate against one another, or as electrons move from one atom to another. Conduction is the most significant means of heat transfer within a solid or between solid objects in thermal contact. Fluids—especially gases—are less conductive. Thermal contact conductance is the study of heat conduction between solid bodies in contact.^[8] The process of heat transfer from one place to another place without the movement of particles is called conduction. Example: Heat transfer through Metal rods. *Steady state conduction* (see Fourier's law) is a form of conduction that happens when the temperature difference driving the conduction is constant, so that after an equilibration time, the spatial distribution of temperatures in the conducting object does not change any further.^[9] In steady state conduction, the amount of heat entering a section is equal to amount of heat coming out.^[8]

Transient conduction (see Heat equation) occurs when the temperature within an object changes as a function of time. Analysis of transient systems is more complex and often calls for the application of approximation theories or numerical analysis by computer.^[8]

Convection

The flow of fluid may be forced by external processes, or sometimes (in gravitational fields) by buoyancy forces caused when thermal energy expands the fluid (for example in a fire plume), thus influencing its own transfer. The latter process is often called "natural convection". All convective processes also move heat partly by diffusion, as well. Another form of convection is forced convection. In this case the fluid is forced to flow by use of a pump, fan or other mechanical means.

Convective heat transfer, or convection, is the transfer of heat from one place to another by the movement of fluids, a process that is essentially the transfer of heat via mass transfer. Bulk motion of fluid enhances heat transfer in many physical situations, such as (for example) between a solid surface and the fluid.^[10] Convection is usually the dominant form of heat transfer in liquids and gases. Although sometimes discussed as a third method of heat transfer, convection is usually used to describe the combined effects of heat conduction within the fluid (diffusion) and heat transference by bulk fluid flow streaming.^[11] The process of transport by fluid streaming is known as advection, but pure advection is a term that is generally associated only with mass transport in fluids, such as advection of pebbles in a river. In the case of heat transfer in fluids, where transport by advection in a fluid is always also accompanied by transport via heat diffusion (also known as heat conduction) the process of heat convection is understood to refer to the sum of heat transport by advection and diffusion/conduction.

Free, or natural, convection occurs when bulk fluid motions (streams and currents) are caused by buoyancy forces that result from density variations due to variations of temperature in the fluid. *Forced* convection is a term used when the streams and currents in the fluid are induced by external means—such as fans, stirrers, and pumps—creating an artificially induced convection current.^[12]

Convection-cooling

Convective cooling is sometimes described as Newton's law of cooling:

The rate of heat loss of a body is proportional to the temperature difference between the body and its surroundings.

However, by definition, the validity of Newton's law of Cooling requires that the rate of heat loss from convection be a linear function of ("proportional to") the temperature difference that drives heat transfer, and in convective cooling this is sometimes not the case. In general, convection is not linearly dependent on temperature gradients, and in some cases is strongly nonlinear. In these cases, Newton's law does not apply.

Convection vs. conduction

In a body of fluid that is heated from underneath its container, conduction and convection can be considered to compete for dominance. If heat conduction is too great, fluid moving down by convection is heated by conduction so fast that its downward movement will be stopped due to its buoyancy, while fluid moving up by convection is cooled by conduction so fast that its driving buoyancy will diminish. On the other hand, if heat conduction is very low, a large temperature gradient may be formed and convection might be very strong.

The Rayleigh number (Ra) is the product of the Grashof and Prandtl numbers. It is a measure which determines the relative strength of conduction and convection.^[13]

$$Ra = GrPr = \frac{g\Delta\rho L^3}{\mu\alpha} = \frac{g\beta\Delta T L^3}{\nu\alpha}$$

where

- g is acceleration due to gravity,
- ρ is the density with $\Delta\rho$ being the density difference between the lower and upper ends,
- μ is the dynamic viscosity,
- α is the Thermal diffusivity,
- β is the volume thermal expansivity (sometimes denoted α elsewhere),
- T is the temperature,
- ν is the kinematic viscosity, and
- L is characteristic length.

The Rayleigh number can be understood as the ratio between the rate of heat transfer by convection to the rate of heat transfer by conduction; or, equivalently, the ratio between the corresponding timescales (i.e. conduction timescale divided by convection timescale), up to a numerical factor. This can be seen as follows, where all calculations are up to numerical factors depending on the geometry of the system.

The buoyancy force driving the convection is roughly $g\Delta\rho L^3$, so the corresponding pressure is roughly $g\Delta\rho L$. In steady state, this is canceled by the shear stress due to viscosity, and therefore roughly equals $\mu V/L = \mu/T_{conv}$, where V is the typical fluid velocity due to convection and T_{conv} the order of its timescale. The conduction timescale, on the other hand, is of the order of $T_{cond} = L^2/\alpha$.

Convection occurs when the Rayleigh number is above 1,000–2,000.

Radiation



Red-hot iron object, transferring heat to the surrounding environment through thermal radiation

Thermal radiation occurs through a vacuum or any transparent medium (solid or fluid). It is the transfer of energy by means of photons in electromagnetic waves governed by the same laws.^[1]

Thermal radiation is energy emitted by matter as electromagnetic waves, due to the pool of thermal energy in all matter with a temperature above absolute zero. Thermal radiation propagates without the presence of matter through the vacuum of space.^[14]

Thermal radiation is a direct result of the random movements of atoms and molecules in matter. Since these atoms and molecules are composed of charged particles (protons and electrons), their movement results in the emission of electromagnetic radiation, which carries energy away from the surface.

The Stefan–Boltzmann equation, which describes the rate of transfer of radiant energy, is as follows for an object in a vacuum :

$$Q = \epsilon\sigma T^4.$$

For radiative transfer between two objects, the equation is as follows:

$$Q = \epsilon\sigma(T_a^4 - T_b^4),$$

where Q is the heat flux, ϵ is the emissivity (unity for a black body), σ is the Stefan–Boltzmann constant, and T is the absolute temperature (in kelvins or degrees Rankine). Radiation is typically only important for very hot objects, or for objects with a large temperature difference.

Radiation from the sun, or solar radiation, can be harvested for heat and power.^[15] Unlike conductive and convective forms of heat transfer, thermal radiation can be concentrated in a small spot by using reflecting mirrors, which is exploited in concentrating solar power generation.^[16] For example, the sunlight reflected from mirrors heats the PS10 solar power tower and during the day it can heat water to 285 °C (545 °F).

Phase transition

Phase transition or phase change, takes place in a thermodynamic system from one phase or state of matter to another one by heat transfer. Phase change examples are the melting of ice or the boiling of water. The Mason equation explains the growth of a water droplet based on the effects of heat transport on evaporation and condensation.

Phase transitions involve the four fundamental states of matter:

- Solid – Deposition, freezing and solid to solid transformation.
- Gas – Boiling/evaporation, recombination/deionization, and sublimation.
- Liquid – Condensation and melting/fusion.
- Plasma – Ionization.

Boiling



Nucleate boiling of water.

The boiling point of a substance is the temperature at which the vapor pressure of the liquid equals the pressure surrounding the liquid^{[18][19]} and the liquid evaporates resulting in an abrupt change in vapor volume.



Lightning is a highly visible form of energy transfer and is an example of plasma present at Earth's surface. Typically, lightning discharges 30,000 amperes at up to 100 million volts, and emits light, radio waves, X-rays and even gamma rays.^[17] Plasma temperatures in lightning can approach 28,000 kelvins (27,726.85 °C) (49,940.33 °F) and electron densities may exceed 10^{24} m^{-3} .

Saturation temperature means boiling point. The saturation temperature is the temperature for a corresponding saturation pressure at which a liquid boils into its vapor phase. The liquid can be said to be saturated with thermal energy. Any addition of thermal energy results in a phase transition.

At standard atmospheric pressure and **low temperatures**, no boiling occurs and the heat transfer rate is controlled by the usual single-phase mechanisms. As the surface temperature is increased, local boiling occurs and vapor bubbles nucleate, grow into the surrounding cooler fluid, and collapse. This is *sub-cooled nucleate boiling*, and is a very efficient heat transfer mechanism. At high bubble generation rates, the bubbles begin to interfere and the heat flux no longer increases rapidly with surface temperature (this is the departure from nucleate boiling, or DNB).

At similar standard atmospheric pressure and **high temperatures**, the hydrodynamically-quieter regime of film boiling is reached. Heat fluxes across the stable vapor layers are low, but rise slowly with temperature. Any contact between fluid and the surface that may be seen probably leads to the extremely rapid nucleation of a fresh vapor layer ("spontaneous nucleation"). At higher temperatures still, a maximum in the heat flux is reached (the critical heat flux, or CHF).

The Leidenfrost Effect demonstrates how nucleate boiling slows heat transfer due to gas bubbles on the heater's surface. As mentioned, gas-phase thermal conductivity is much lower than liquid-phase thermal conductivity, so the outcome is a kind of "gas thermal barrier".

Condensation

Condensation occurs when a vapor is cooled and changes its phase to a liquid. During condensation, the latent heat of vaporization must be released. The amount of the heat is the same as that absorbed during vaporization at the same fluid pressure.^[20]

There are several types of condensation:

- Homogeneous condensation, as during a formation of fog.
- Condensation in direct contact with subcooled liquid.
- Condensation on direct contact with a cooling wall of a heat exchanger: This is the most common mode used in industry:
 - Filmwise condensation is when a liquid film is formed on the subcooled surface, and usually occurs when the liquid wets the surface.
 - Dropwise condensation is when liquid drops are formed on the subcooled surface, and usually occurs when the liquid does not wet the surface.

Dropwise condensation is difficult to sustain reliably; therefore, industrial equipment is normally designed to operate in filmwise condensation mode.

Melting

Melting is a thermal process that results in the phase transition of a substance from a solid to a liquid. The internal energy of a substance is increased, typically with heat or pressure, resulting in a rise of its temperature to the melting point, at which the ordering of ionic or molecular entities in the solid breaks down to a less ordered state and the solid liquefies. Molten substances generally have reduced viscosity with elevated temperature; an exception to this maxim is the element sulfur, whose viscosity increases to a point due to polymerization and then decreases with higher temperatures in its molten state.^[21]

Modeling approaches

Heat transfer can be modeled in the following ways.

Climate models

Climate models study the radiant heat transfer by using quantitative methods to simulate the interactions of the atmosphere, oceans, land surface, and ice.

Heat equation

The heat equation is an important partial differential equation that describes the distribution of heat (or variation in temperature) in a given region over time. In some cases, exact solutions of the equation are available;^[22] in other cases the equation must be solved numerically using computational methods.

Lumped system analysis

Lumped system analysis often reduces the complexity of the equations to one first-order linear differential equation, in which case heating and cooling are described by a simple exponential solution, often referred to as Newton's law of cooling.

System analysis by the lumped capacitance model is a common approximation in transient conduction that may be used whenever heat conduction within an object is much faster than heat conduction across the boundary of the object. This is a method of approximation that reduces one aspect of the transient conduction system—that within the object—to an equivalent steady state system. That is, the method assumes that the temperature within the object is completely uniform, although its value may be changing in time.



Ice melting

In this method, the ratio of the conductive heat resistance within the object to the convective heat transfer resistance across the object's boundary, known as the *Biot number*, is calculated. For small Biot numbers, the approximation of *spatially uniform temperature within the object* can be used: it can be presumed that heat transferred into the object has time to uniformly distribute itself, due to the lower resistance to doing so, as compared with the resistance to heat entering the object.^[23]

Engineering

Heat transfer has broad application to the functioning of numerous devices and systems. Heat-transfer principles may be used to preserve, increase, or decrease temperature in a wide variety of circumstances. Heat transfer methods are used in numerous disciplines, such as automotive engineering, thermal management of electronic devices and systems, climate control, insulation, materials processing, and power station engineering.



Heat exposure as part of a fire test for firestop products

Insulation, radiance and resistance

Thermal insulators are materials specifically designed to reduce the flow of heat by limiting conduction, convection, or both. Thermal resistance is a heat property and the measurement by which an object or material resists to heat flow (heat per time unit or thermal resistance) to temperature difference.

Radiance or spectral radiance are measures of the quantity of radiation that passes through or is emitted. Radiant barriers are materials that reflect radiation, and therefore reduce the flow of heat from radiation sources. Good insulators are not necessarily good radiant barriers, and vice versa. Metal, for instance, is an excellent reflector and a poor insulator.

The effectiveness of a radiant barrier is indicated by its **reflectivity**, which is the fraction of radiation reflected. A material with a high reflectivity (at a given wavelength) has a low emissivity (at that same wavelength), and vice versa. At any specific wavelength, $\text{reflectivity} = 1 - \text{emissivity}$. An ideal radiant barrier would have a reflectivity of 1, and would therefore reflect 100 percent of incoming radiation. Vacuum flasks, or Dewars, are silvered to approach this ideal. In the vacuum of space, satellites use multi-layer insulation, which consists of many layers of aluminized (shiny) Mylar to greatly reduce radiation heat transfer and control satellite temperature.

Devices

- Heat engine is a system that performs the conversion of heat or thermal energy to mechanical energy which can then be used to do mechanical work.^{[24][25]}
- Thermocouple is a temperature-measuring device and widely used type of temperature sensor for measurement and control, and can also be used to convert heat into electric power.
- Thermoelectric cooler is a solid state electronic device that pumps (transfers) heat from one side of the device to the other when electric current is passed through it. It is based on the Peltier effect.
- Thermal diode or thermal rectifier is a device that causes heat to flow preferentially in one direction.

Heat exchangers

A heat exchanger is used for more efficient heat transfer or to dissipate heat. Heat exchangers are widely used in refrigeration, air conditioning, space heating, power generation, and chemical processing. One common example of a heat exchanger is a car's radiator, in which the hot coolant fluid is cooled by the flow of air over the radiator's surface.

Common types of heat exchanger flows include parallel flow, counter flow, and cross flow. In parallel flow, both fluids move in the same direction while transferring heat; in counter flow, the fluids move in opposite directions; and in cross flow, the fluids move at right angles to each other. Common constructions for heat exchanger include shell and tube, double pipe, extruded finned pipe, spiral fin pipe, u-tube, and stacked plate.

A heat sink is a component that transfers heat generated within a solid material to a fluid medium, such as air or a liquid. Examples of heat sinks are the heat exchangers used in refrigeration and air conditioning systems or the radiator in a car. A heat pipe is another heat-transfer device that combines thermal conductivity and phase transition to efficiently transfer heat between two solid interfaces.

Examples

Architecture

Efficient energy use is the goal to reduce the amount of energy required in heating or cooling. In architecture, condensation and air currents can cause cosmetic or structural damage. An energy audit can help to assess the implementation of recommended corrective procedures. For instance, insulation improvements, air sealing of structural leaks or the addition of energy-efficient windows and doors.^[26]

- Smart meter is a device that records electric energy consumption in intervals.
- Thermal transmittance is the rate of transfer of heat through a structure divided by the difference in temperature across the structure. It is expressed in watts per square meter per kelvin, or $W/(m^2K)$. Well-insulated parts of a building have a low thermal transmittance, whereas poorly-insulated parts of a building have a high thermal transmittance.
- Thermostat is a device to monitor and control temperature.

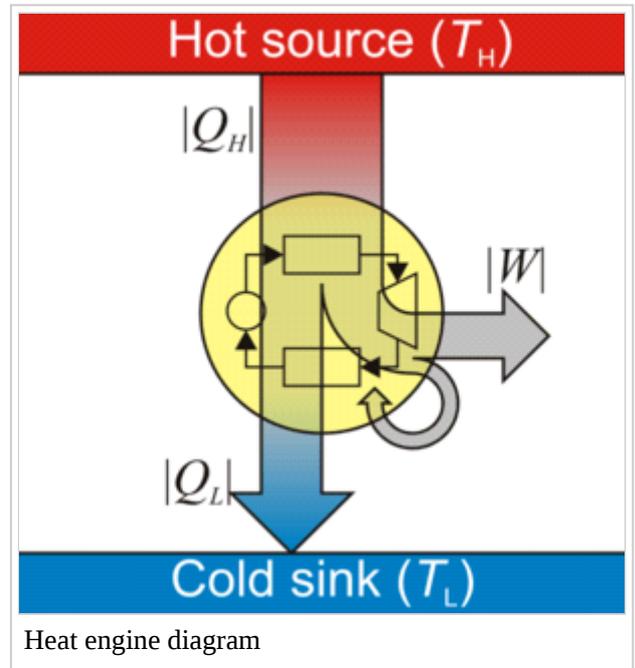
Climate engineering

Climate engineering consists of carbon dioxide removal and solar radiation management. Since the amount of carbon dioxide determines the radiative balance of Earth atmosphere, carbon dioxide removal techniques can be applied to reduce the radiative forcing. Solar radiation management is the attempt to absorb less solar radiation to offset the effects of greenhouse gases.

Greenhouse effect

The greenhouse effect is a process by which thermal radiation from a planetary surface is absorbed by atmospheric greenhouse gases, and is re-radiated in all directions. Since part of this re-radiation is back towards the surface and the lower atmosphere, it results in an elevation of the average surface temperature above what it would be in the absence of the gases.

Heat transfer in the human body



An example application in climate engineering includes the creation of Biochar through the pyrolysis process. Thus, storing greenhouse gases in carbon reduces the radiative forcing capacity in the atmosphere, causing more long-wave (infrared) radiation out to Space.

The principles of heat transfer in engineering systems can be applied to the human body in order to determine how the body transfers heat. Heat is produced in the body by the continuous metabolism of nutrients which provides energy for the systems of the body.^[27] The human body must maintain a consistent internal temperature in order to maintain healthy bodily functions. Therefore, excess heat must be dissipated from the body to keep it from overheating. When a person engages in elevated levels of physical activity, the body requires additional fuel which increases the metabolic rate and the rate of heat production. The body must then use additional methods to remove the additional heat produced in order to keep the internal temperature at a healthy level.

Heat transfer by convection is driven by the movement of fluids over the surface of the body. This convective fluid can be either a liquid or a gas. For heat transfer from the outer surface of the body, the convection mechanism is dependent on the surface area of the body, the velocity of the air, and the temperature gradient between the surface of the skin and the ambient air.^[28] The normal temperature of the body is approximately 37 °C. Heat transfer occurs more readily when the temperature of the surroundings is significantly less than the normal body temperature. This concept explains why a person feels “cold” when not enough covering is worn when exposed to a cold environment. Clothing can be considered an insulator which provides thermal resistance to heat flow over the covered portion of the body.^[29] This thermal resistance causes the temperature on the surface of the clothing to be less than the temperature on the surface of the skin. This smaller temperature gradient between the surface temperature and the ambient temperature will cause a lower rate of heat transfer than if the skin were not covered.

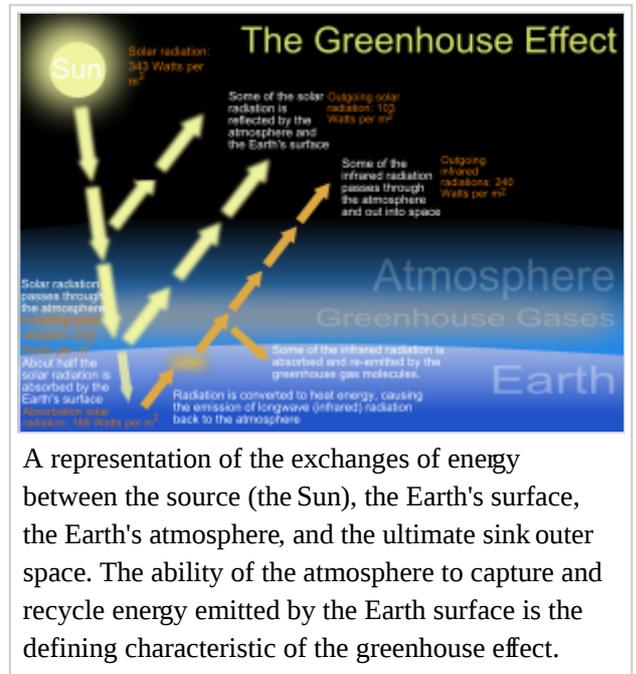
In order to ensure that one portion of the body is not significantly hotter than another portion, heat must be distributed evenly through the bodily tissues. Blood flowing through blood vessels acts as a convective fluid and helps to prevent any buildup of excess heat inside the tissues of the body. This flow of blood through the vessels can be modeled as pipe flow in an engineering system. The heat carried by the blood is determined by the temperature of the surrounding tissue, the diameter of the blood vessel, the thickness of the fluid, velocity of the flow, and the heat transfer coefficient of the blood. The velocity, blood vessel diameter, and the fluid thickness can all be related with the Reynolds Number, a dimensionless number used in fluid mechanics to characterize the flow of fluids.

Latent heat loss, also known as evaporative heat loss, accounts for a large fraction of heat loss from the body. When the core temperature of the body increases, the body triggers sweat glands in the skin to bring additional moisture to the surface of the skin. The liquid is then transformed into vapor which removes heat from the surface of the body.^[30] The rate of evaporation heat loss is directly related to the vapor pressure at the skin surface and the amount of moisture present on the skin.^[28] Therefore, the maximum of heat transfer will occur when the skin is completely wet. The body continuously loses water by evaporation but the most significant amount of heat loss occurs during periods of increased physical activity.

Cooling techniques

Evaporative cooling

Evaporative cooling happens when water vapor is added to the surrounding air. The energy needed to evaporate the water is taken from the air in the form of sensible heat and converted into latent heat, while the air remains at a constant enthalpy. Latent heat describes the amount of heat that is needed to evaporate the liquid; this heat comes from the liquid itself and the surrounding gas and surfaces. The greater the difference between the two



A representation of the exchanges of energy between the source (the Sun), the Earth's surface, the Earth's atmosphere, and the ultimate sink outer space. The ability of the atmosphere to capture and recycle energy emitted by the Earth surface is the defining characteristic of the greenhouse effect.

temperatures, the greater the evaporative cooling effect. When the temperatures are the same, no net evaporation of water in air occurs; thus, there is no cooling effect.

Laser cooling

In Quantum Physics laser cooling is used to achieve temperatures of near absolute zero ($-273.15\text{ }^{\circ}\text{C}$, $-459.67\text{ }^{\circ}\text{F}$) of atomic and molecular samples, to observe unique quantum effects that can only occur at this heat level.

- Doppler cooling is the most common method of laser cooling.
- Sympathetic cooling is a process in which particles of one type cool particles of another type. Typically, atomic ions that can be directly laser-cooled are used to cool nearby ions or atoms. This technique allows cooling of ions and atoms that cannot be laser cooled directly.

Magnetic cooling

Magnetic evaporative cooling is a process for lowering the temperature of a group of atoms, after pre-cooled by methods such as laser cooling. Magnetic refrigeration cools below 0.3K , by making use of the magnetocaloric effect.

Radiative cooling

Radiative cooling is the process by which a body loses heat by radiation. Outgoing energy is an important effect in the Earth's energy budget. In the case of the Earth-atmosphere system, it refers to the process by which long-wave (infrared) radiation is emitted to balance the absorption of short-wave (visible) energy from the Sun. Convective transport of heat and evaporative transport of latent heat both remove heat from the surface and redistribute it in the atmosphere.

Thermal energy storage

Thermal energy storage includes technologies for collecting and storing energy for later use. It may be employed to balance energy demand between day and nighttime. The thermal reservoir may be maintained at a temperature above or below that of the ambient environment. Applications include space heating, domestic or process hot water systems, or generating electricity.

See also

- Combined forced and natural convection
- Heat capacity
- Heat transfer physics
- Stefan–Boltzmann law
- Thermal contact conductance
- Thermal physics
- Thermal resistance in electronics
- Thermal science
- Heat transfer enhancement

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External links

- A Heat Transfer Textbook - (free download).
- Thermal-FluidsPedia - An online thermal fluids encyclopedia.
- Hyperphysics Article on Heat Transfer - Overview
- Interseasonal Heat Transfer - a practical example of how heat transfer is used to heat buildings without burning fossil fuels.
- Aspects of Heat Transfer, Cambridge University
- Thermal-Fluids Central
- Energy2D: Interactive Heat Transfer Simulations for Everyone

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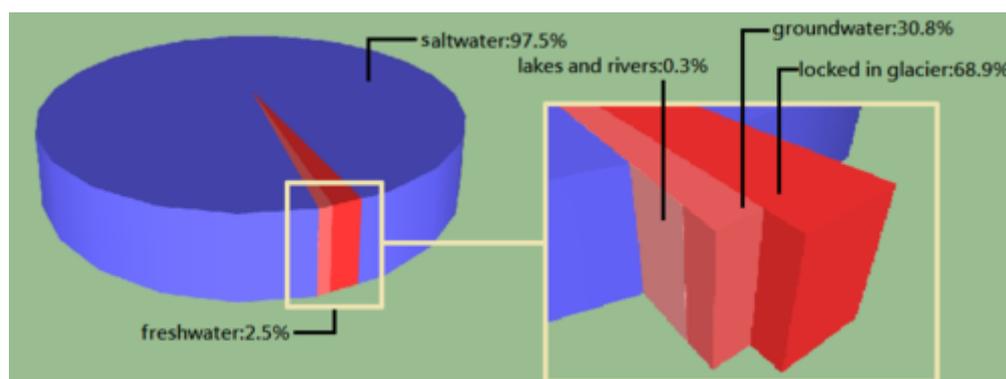
Hydrosphere

From Wikipedia, the free encyclopedia

The **hydrosphere** (from Greek ὕδωρ *hydōr*, "water"^[1] and σφαῖρα *sphaira*, "sphere"^[2]) is the combined mass of water found on, under, and above the surface of a planet, minor planet or natural satellite.

It has been estimated that there are 1386 million cubic

kilometers of water on Earth.^[3] This includes water in liquid and frozen forms in groundwater, oceans, lakes and streams. Saltwater accounts for 97.5% of this amount. Fresh water accounts for only 2.5%. Of this fresh water, 68.9% is in the form of ice and permanent snow cover in the Arctic, the Antarctic, and mountain glaciers. 30.8% is in the form of fresh groundwater. Only 0.3% of the fresh water on Earth is in easily accessible lakes, reservoirs and river systems.^[3] The total mass of the Earth's hydrosphere is about 1.4×10^{18} tonnes, which is about 0.023% of Earth's total mass. About 20×10^{12} tonnes of this is in Earth's atmosphere (for practical purposes, 1 cubic metre of water weighs one tonne). Approximately 75% of Earth's surface, an area of some 361 million square kilometers (139.5 million square miles), is covered by ocean. The average salinity of Earth's oceans is about 35 grams of salt per kilogram of sea water (3.5%).^[4]



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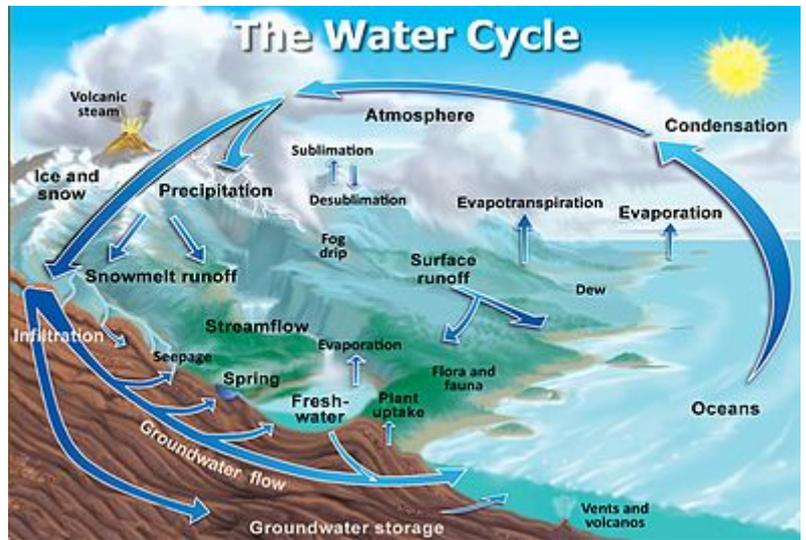
Water cycle

The hydrological cycle transfers water from one state or reservoir to another. Reservoirs include atmospheric moisture (snow, rain and clouds), streams, oceans, rivers, lakes, groundwater, subterranean aquifers, polar icecaps and saturated soil. Solar energy, in the form of heat and light (insolation), and gravity cause the transfer from one state to another over periods from hours to thousands of years. Most evaporation comes from the oceans and is returned to the earth as snow or rain (page 27).^[5] Sublimation refers to evaporation from snow and ice. Transpiration refers to the expiration of water through the minute pores or stomata of trees. Evapotranspiration is the term used by hydrologists in reference to the three processes together, transpiration, sublimation and evaporation.^[5]

In his book *Water*, Marq de Villiers described the hydrosphere as a closed system in which water exists. The hydrosphere is intricate, complex, interdependent, all-pervading and stable and "seems purpose-built for regulating life (de Villiers 2003:26)."^[5] De Villiers claimed that, "On earth, the total amount of water has almost certainly not changed since geological times: what we had then we still have. Water can be polluted,

abused, and misused but it is neither created nor destroyed, it only migrates. There is no evidence that water vapor escapes into space (page 26)."^[5]

"Every year the turnover of water on Earth involves 577,000 km³ of water. This is water that evaporates from the oceanic surface (502,800 km³) and from land (74,200 km³). The same amount of water falls as atmospheric precipitation, 458,000 km³ on the ocean and 119,000 km³ on land. The difference between precipitation and evaporation from the land surface (119,000 - 74,200 = 44,800 km³/year) represents the total runoff of the Earth's rivers (42,700 km³/year) and direct groundwater runoff to the ocean (2100 km³/year). These are the principal sources of fresh water to support life necessities and man's economic activities."^[3]



Water is a basic necessity of life. Since 2/3 of the Earth is covered by water, the Earth is also called the blue planet and the watery planet.^[notes 1] Hydrosphere plays an important role in the existence of the atmosphere in its present form. Oceans are important in this regard. When the Earth was formed it had only a very thin atmosphere rich in hydrogen and helium similar to the present atmosphere of Mercury. Later the gases hydrogen and helium were expelled from the atmosphere. The gases and water vapor released as the Earth cooled became our present atmosphere. Other gases and water vapor released by volcanoes also entered the atmosphere. As the Earth cooled the water vapor in the atmosphere condensed and fell as rain. The atmosphere cooled further as atmospheric carbon dioxide dissolved in to rain water. In turn this further caused the water vapor to condense and fall as rain. This rain water filled the depressions on the Earth's surface and formed the oceans. It is estimated that this occurred about 4000 million years ago. The first life forms began in the oceans. These organisms did not breathe oxygen. Later, when cyanobacteria evolved, the process of conversion of carbon dioxide into food and oxygen began. As a result, our atmosphere has a distinctly different composition from that of the other planets; it is a fundamental requirement for life on Earth.

Recharging reservoirs

According to Igor A. Shiklomanov, it takes 2500 years for the complete recharge and replenishment of oceanic waters, 10,000 years for permafrost and ice, 1500 years for deep groundwater and mountainous glaciers, 17 years in lakes and 16 days in rivers.^[3]

Specific fresh water availability

"Specific water availability is the residual (after use) per capita quantity of fresh water."^[3] Fresh water resources are unevenly distributed in terms of space and time and can go from floods to water shortages within months in the same area. In 1998 76% of the total population had a specific water availability of less than 5.0 thousand m³ per year per capita. Already by 1998, 35% of the global population suffered "very low or catastrophically low water supplies" and Shiklomanov predicted that the situation would deteriorate in the twenty-first century with "most of the Earth's population will be living under the conditions of low or catastrophically low water supply" by 2025. There is only 2.5% of fresh water in the hydrosphere.

See also

- Biosphere

- Cryosphere
- Earth's atmosphere
- Lithosphere
- Ocean
- Pedosphere
- Water cycle

Notes

1. According to planetary geologist, Ronald Greeley, "Water is very common in the outer solar system." Europa holds more water than earth's oceans.

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External links

- Ground Water - USGS

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Inverse-square law

From Wikipedia, the free encyclopedia

In physics, an **inverse-square law** is any physical law stating that a specified physical quantity or intensity is inversely proportional to the square of the distance from the source of that physical quantity. The fundamental cause for this can be understood as geometric dilution corresponding to point-source radiation into three-dimensional space (see diagram). Mathematically formulated:

$$\text{intensity} \propto \frac{1}{\text{distance}^2}$$

It can also be mathematically expressed as:

$$\frac{\text{intensity}_1}{\text{intensity}_2} = \frac{\text{distance}_2^2}{\text{distance}_1^2}$$

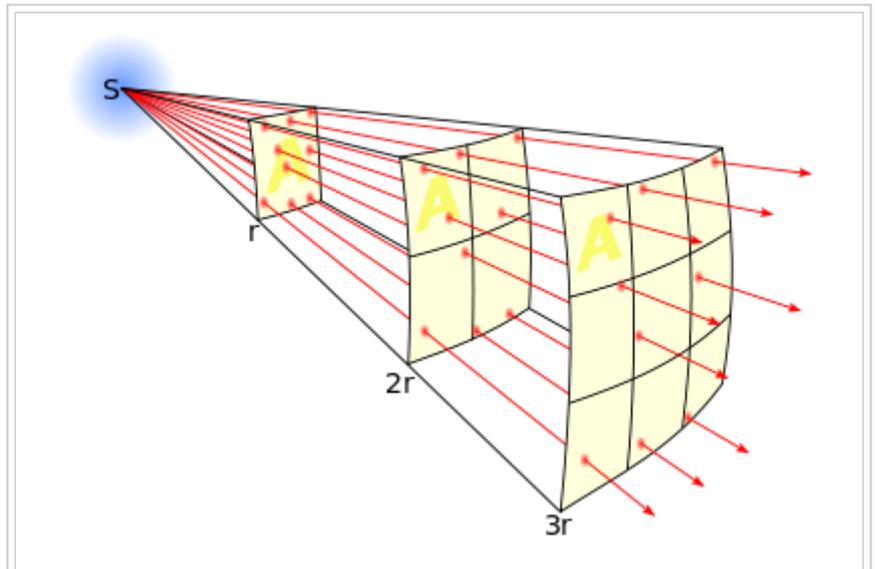
or as the formulation of a constant quantity:

$$\text{intensity}_1 \times \text{distance}_1^2 = \text{intensity}_2 \times \text{distance}_2^2$$

The divergence of a vector field which is the resultant of radial inverse-square law fields with respect to one or more sources is everywhere proportional to the strength of the local sources, and hence zero outside sources. Newton's law of universal gravitation follows an inverse-square law, as do the effects of electric, magnetic, light, sound, and radiation phenomena.

Radar energy expands during both the signal transmission and also on the reflected return, so the inverse square for both paths means that the radar will receive energy according to the inverse fourth power of the range.

In order to prevent dilution of energy while propagating a signal, certain methods can be used such as a waveguide, which acts like a canal does for water, or how a gun barrel restricts hot gas expansion to one dimension in order to prevent loss of energy transfer to a bullet.



S represents the light source, while r represents the measured points. The lines represent the flux emanating from the source. The total number of flux lines depends on the strength of the source and is constant with increasing distance. A greater density of flux lines (lines per unit area) means a stronger field. The density of flux lines is inversely proportional to the square of the distance from the source because the surface area of a sphere increases with the square of the radius. Thus the strength of the field is inversely proportional to the square of the distance from the source.

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Justification

The inverse-square law generally applies when some force, energy, or other conserved quantity is evenly radiated outward from a point source in three-dimensional space. Since the surface area of a sphere (which is $4\pi r^2$) is proportional to the square of the radius, as the emitted radiation gets farther from the source, it is spread out over an area that is increasing in proportion to the square of the distance from the source. Hence, the intensity of radiation passing through any unit area (directly facing the point source) is inversely proportional to the square of the distance from the point source. Gauss's law is similarly applicable, and can be used with any physical quantity that acts in accord to the inverse-square relationship.

Occurrences

Gravitation

Gravitation is the attraction of two objects with mass. Newton's law states:

*The gravitational attraction force between two **point masses** is directly proportional to the product of their masses and inversely proportional to the square of their separation distance. The force is always attractive and acts along the line joining them.*

If the distribution of matter in each body is spherically symmetric, then the objects can be treated as point masses without approximation, as shown in the shell theorem. Otherwise, if we want to calculate the attraction between massive bodies, we need to add all the point-point attraction forces vectorially and the net attraction might not be exact inverse square. However, if the separation between the massive bodies is much larger compared to their sizes, then to a good approximation, it is reasonable to treat the masses as point mass while calculating the gravitational force.

As the law of gravitation, this law was suggested in 1645 by Ismael Bullialdus. But Bullialdus did not accept Kepler's second and third laws, nor did he appreciate Christiaan Huygens's solution for circular motion (motion in a straight line pulled aside by the central force). Indeed, Bullialdus maintained the sun's force was attractive at aphelion and repulsive at perihelion. Robert Hooke and Giovanni Alfonso Borelli both expounded gravitation in 1666 as an attractive force^[1] (Hooke's lecture "On gravity" at the Royal Society, London, on 21 March;^[2] Borelli's "Theory of the Planets", published later in 1666^[3]). Hooke's 1670 Gresham lecture explained that gravitation applied to "all celestiall bodys" and added the principles that the gravitating power decreases with distance and that in the absence of any such power bodies move in straight lines. By 1679, Hooke thought gravitation had inverse square dependence and communicated this in a letter to Isaac Newton.^[4] Hooke remained bitter about Newton claiming the invention of this principle, even though Newton's *Principia* acknowledged that Hooke, along with Wren and Halley, had separately appreciated the inverse square law in the solar system,^[5] as well as giving some credit to Bullialdus.^[6]

Electrostatics

The force of attraction or repulsion between two electrically charged particles, in addition to being directly proportional to the product of the electric charges, is inversely proportional to the square of the distance between them; this is known as Coulomb's law. The deviation of the exponent from 2 is less than one part in 10^{15} .^[7]

Light and other electromagnetic radiation

The intensity (or illuminance or irradiance) of light or other linear waves radiating from a point source (energy per unit of area perpendicular to the source) is inversely proportional to the square of the distance from the source; so an object (of the same size) twice as far away, receives only one-quarter the energy (in the same time period).

More generally, the irradiance, *i.e.*, the intensity (or power per unit area in the direction of propagation), of a spherical wavefront varies inversely with the square of the distance from the source (assuming there are no losses caused by absorption or scattering).

For example, the intensity of radiation from the Sun is 9126 watts per square meter at the distance of Mercury (0.387 AU); but only 1367 watts per square meter at the distance of Earth (1 AU)—an approximate threefold increase in distance results in an approximate ninefold decrease in intensity of radiation.

For non-isotropic radiators such as parabolic antennas, headlights, and lasers, the effective origin is located far behind the beam aperture. If you are close to the origin, you don't have to go far to double the radius, so the signal drops quickly. When you are far from the origin and still have a strong signal, like with a laser, you have to travel very far to double the radius and reduce the signal. This means you have a stronger signal or have antenna gain in the direction of the narrow beam relative to a wide beam in all directions of an isotropic antenna.

In photography and stage lighting, the inverse-square law is used to determine the "fall off" or the difference in illumination on a subject as it moves closer to or further from the light source. For quick approximations, it is enough to remember that doubling the distance reduces illumination to one quarter;^[8] or similarly, to halve the illumination increase the distance by a factor of 1.4 (the square root of 2), and to double illumination, reduce the distance to 0.7 (square root of 1/2). When the illuminant is not a point source, the inverse square rule is often still a useful approximation; when the size of the light source is less than one-fifth of the distance to the subject, the calculation error is less than 1%.^[9]

The fractional reduction in electromagnetic fluence (Φ) for indirectly ionizing radiation with increasing distance from a point source can be calculated using the inverse-square law. Since emissions from a point source have radial directions, they intercept at a perpendicular incidence. The area of such a shell is $4\pi r^2$ where r is the radial distance from the center. The law is particularly important in diagnostic radiography and radiotherapy treatment planning, though this proportionality does not hold in practical situations unless source dimensions are much smaller than the distance. As stated in Fourier theory of heat "as the point source is magnification by distances, its radiation is dilute proportional to the sin of the angle, of the increasing circumference arc from the point of origin"

Example

Let the total power radiated from a point source, for example, an omnidirectional isotropic radiator, be P . At large distances from the source (compared to the size of the source), this power is distributed over larger and larger spherical surfaces as the distance from the source increases. Since the surface area of a sphere of radius r is $A = 4\pi r^2$, then intensity I (power per unit area) of radiation at distance r is

$$I = \frac{P}{A} = \frac{P}{4\pi r^2}.$$

The energy or intensity decreases (divided by 4) as the distance r is doubled; measured in dB it would decrease by 6.02 dB per doubling of distance.

Sound in a gas

In acoustics, the sound pressure of a spherical wavefront radiating from a point source decreases by 50% as the distance r is doubled; measured in dB, the decrease is still 6.02 dB, since dB represents an intensity ratio. The pressure ratio (as opposed to power ratio) is not inverse-square, but is inverse-proportional (inverse distance law):

$$p \propto \frac{1}{r}$$

The same is true for the component of particle velocity v that is in-phase with the instantaneous sound pressure p :

$$v \propto \frac{1}{r}$$

In the near field is a quadrature component of the particle velocity that is 90° out of phase with the sound pressure and does not contribute to the time-averaged energy or the intensity of the sound. The sound intensity is the product of the RMS sound pressure and the *in-phase* component of the RMS particle velocity, both of which are inverse-proportional. Accordingly, the intensity follows an inverse-square behaviour:

$$I = pv \propto \frac{1}{r^2}.$$

Field theory interpretation

For an irrotational vector field in three-dimensional space, the inverse-square law corresponds to the property that the divergence is zero outside the source. This can be generalized to higher dimensions. Generally, for an irrotational vector field in n -dimensional Euclidean space, the intensity "I" of the vector field falls off with the distance "r" following the inverse $(n - 1)$ th power law

$$I \propto \frac{1}{r^{n-1}},$$

given that the space outside the source is divergence free.

History

John Dumbleton of the 14th-century Oxford Calculators, was one of the first to express functional relationships in graphical form. He gave a proof of the mean speed theorem stating that "the latitude of a uniformly difform movement corresponds to the degree of the midpoint" and used this method to study the quantitative decrease in intensity of illumination in his *Summa logicæ et philosophiæ naturalis* (ca. 1349), stating that it was not linearly proportional to the distance, but was unable to expose the Inverse-square law.^[10]

In proposition 9 of Book 1 in his book *Ad Vitellionem paralipomena, quibus astronomiæ pars optica traditur* (1604), the astronomer Johannes Kepler argued that the spreading of light from a point source obeys an inverse square law:^{[11][12]}

Original: *Sicut se habent sphaericae superficies, quibus origo lucis pro centro est, amplior ad angustiores: ita se habet fortitudo seu densitas lucis radorum in angustiori, ad illamin in laxiori sphaerica, hoc est, conversim. Nam per 6. 7. tantundem lucis est in angustiori sphaerica superficie, quantum in fusiore, tanto ergo illie stipatior & densior quam hic.*

Translation: Just as [the ratio of] spherical surfaces, for which the source of light is the center, [is] from the wider to the narrower, so the density or fortitude of the rays of light in the narrower [space], towards the more spacious spherical surfaces, that is, inversely. For according to [propositions] 6 & 7, there is as much light in the narrower spherical surface, as in the wider, thus it is as much more compressed and dense here than there.

In 1645 in his book *Astronomia Philolaica ...*, the French astronomer Ismaël Bullialdus (1605 – 1694) refuted Johannes Kepler's suggestion that "gravity"^[13] weakens as the inverse of the distance; instead, Bullialdus argued, "gravity" weakens as the inverse square of the distance:^{[14][15]}

Original: Virtus autem illa, qua Solprehendit seu harpagat planetas, corporalis quae ipsi pro manibus est, lineis rectis in omnem mundi amplitudinem emissa quasi species solis cum illius corpore rotatur: cum ergo sit corporalis imminuitur, & extenuatur in maiori spatio & intervallo, ratio autem huius imminutionis eadem est, ac luminum, in ratione nempe dupla intervallorum, sed eversa.

Translation: As for the power by which the Sun seizes or holds the planets, and which, being corporeal, functions in the manner of hands, it is emitted in straight lines throughout the whole extent of the world, and like the species of the Sun, it turns with the body of the Sun; now, seeing that it is corporeal, it becomes weaker and attenuated at a greater distance or interval, and the ratio of its decrease in strength is the same as in the case of light, namely, the duplicate proportion, but inversely, of the distances [that is, $1/d^2$].

In England, the Anglican bishop Seth Ward (1617 – 1689) publicized the ideas of Bullialdus in his critique *In Ismaelis Bullialdi astronomiae philolaicae fundamenta inquisitio brevis* (1653) and publicized the planetary astronomy of Kepler in his book *Astronomia geometrica* (1656).

In 1663–1664, the English scientist Robert Hooke was writing his book *Micrographia* (1666) in which he discussed, among other things, the relation between the height of the atmosphere and the barometric pressure at the surface. Since the atmosphere surrounds the earth, which itself is a sphere, the volume of atmosphere bearing on any unit area of the earth's surface is a truncated cone (which extends from the earth's center to the vacuum of space; obviously only the section of the cone from the earth's surface to space bears on the earth's surface). Although the volume of a cone is proportional to the cube of its height, Hooke argued that the air's pressure at the earth's surface is instead proportional to the height of the atmosphere because gravity diminishes with altitude. Although Hooke did not explicitly state so, the relation that he proposed would be true only if gravity decreases as the inverse square of the distance from the earth's center.^{[16][17]}

See also

- Flux
- Gauss's law
- Kepler's first law
- Telecommunications, particularly:
 - William Thomson, 1st Baron Kelvin
 - Power-aware routing protocols
- Inverse proportionality
- Multiplicative inverse
- Distance decay

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13. Note: Kepler's notion of "gravity" was not the modern notion: Kepler believed that the Sun exerted a force on the planets which propelled them in their orbits around the Sun. Like beams of light from a light house, this force rotated with the Sun as it turned.
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16. (Gal & Chen-Morris, 2005), pp. 391–392.
17. Robert Hooke, *Micrographia...* (London, England: John Martyn, 1667) page 227: (<http://digicoll.librarywisc.edu/cgi-bin/HistSciTech/HistSciTech-idx?type=goto&id=HistSciTech.HookeMicro&isize=M&submit=Go+to+page&page=227>) "[I say a *Cylinder*, not a piece of a *Cone*, because, as I may elsewhere shew in the Explication of Gravity, that *triplicate* proportion of the shells of a Sphere, to their respective diameters, I suppose to be removed in this case by the decrease the power of Gravity]"

External links

- Damping of sound level with distance
- Sound pressure p and the inverse distance law 1/r
- Inverse Square Law & Radiation Protection by Ionactive (Animation)

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Lapse rate

From Wikipedia, the free encyclopedia

The **lapse rate** is the rate at which atmospheric temperature decreases with an increase in altitude.^{[1][2]} The terminology arises from the word *lapse* in the sense of a decrease or decline. While most often applied to Earth's troposphere, the concept can be extended to any gravitationally supported parcel of gas.

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Definition

A formal definition from the *Glossary of Meteorology*^[3] is:

The decrease of an atmospheric variable with height, the variable being temperature unless otherwise specified.

In general, a lapse rate is the negative of the rate of temperature change with altitude change, thus:

$$\gamma = -\frac{dT}{dz}$$

where γ is the lapse rate given in units of temperature divided by units of altitude, T = temperature, and z = altitude.^[a]

Convection and adiabatic expansion

The temperature profile of the atmosphere is a result of an interaction between radiation and convection. Sunlight hits the ground and heats it. The ground then heats the air at the surface. If radiation were the only way to transfer heat from the ground to space, the greenhouse effect of gases in the atmosphere would keep the ground at roughly 333 K (60 °C; 140 °F), and the temperature would decay exponentially with height.^[6]

However, when air is hot, it tends to expand, which lowers its density. Thus, hot air tends to rise and transfer heat upward. This is the process of convection. Convection comes to equilibrium when a parcel of air at a given altitude has the same density as the other air at the same elevation.

When a parcel of air expands, it pushes on the air around it, doing work (thermodynamics). Since the parcel does work but gains no heat, it loses internal energy so that its temperature decreases. The process of expanding and contracting without exchanging heat is an adiabatic process. The term *adiabatic* means that no heat transfer

occurs into or out of the parcel. Air has low thermal conductivity, and the bodies of air involved are very large, so transfer of heat by conduction is negligibly small.

The adiabatic process for air has a characteristic temperature-pressure curve, so the process determines the lapse rate. When the air contains little water, this lapse rate is known as the dry adiabatic lapse rate: the rate of temperature decrease is 9.8 °C/km (5.38 °F per 1,000 ft) (3.0 °C/1,000 ft). The reverse occurs for a sinking parcel of air.^[7]

Note that only the troposphere (up to approximately 12 kilometres (39,000 ft) of altitude) in the Earth's atmosphere undergoes convection: the stratosphere does not generally convect.^[8] However, some exceptionally energetic convection processes -- notably volcanic eruption columns and overshooting tops associated with severe supercell thunderstorms -- may *locally* and *temporarily* inject convection through the tropopause and into the stratosphere.

The mathematics of the adiabatic lapse rate can be derived from thermodynamics, which defines an adiabatic process via:

$$PdV = -VdP/\gamma$$

the first law of thermodynamics can be written as

$$mc_v dT - Vdp/\gamma = 0$$

Also since $\alpha = V/m$ and $\gamma = c_p/c_v$ we can show that:

$$c_p dT - \alpha dP = 0$$

where c_p is the specific heat at constant pressure and α is the specific volume.

Assuming an atmosphere in hydrostatic equilibrium:^[9]

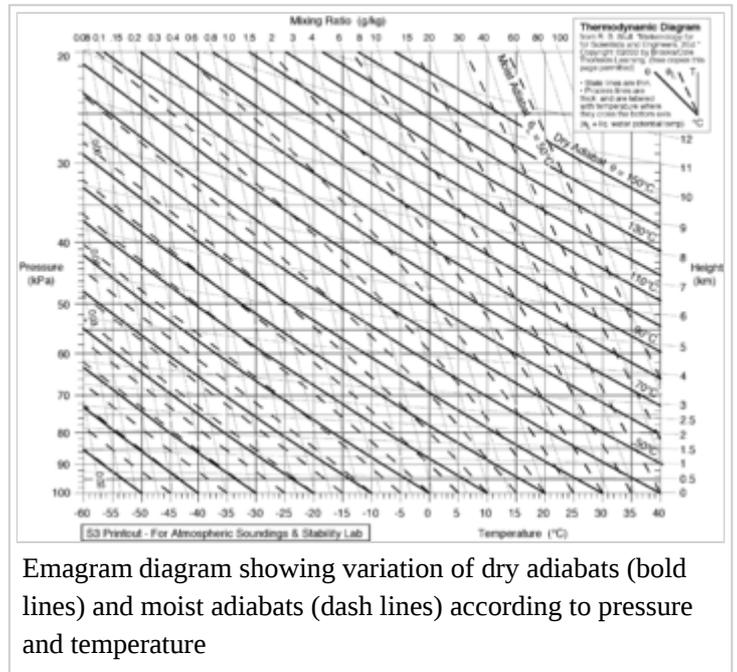
$$dP = -\rho g dz$$

where g is the standard gravity and ρ is the density. Combining these two equations to eliminate the pressure, one arrives at the result for the dry adiabatic lapse rate (DALR),^[10]

$$\Gamma_d = -\frac{dT}{dz} = \frac{g}{c_p} = 9.8 \text{ °C/km}$$

Moist adiabatic lapse rate

The presence of water within the atmosphere complicates the process of convection. Water vapor contains latent heat of vaporization. As air rises and cools, it eventually becomes saturated and cannot hold its quantity of water vapor. The water vapor condenses, forming clouds, and releasing heat. Before saturation, the rising air follows the dry adiabatic lapse rate. After saturation, the rising air follows the moist adiabatic lapse rate.^[11] The release of latent heat is an important source of energy in the development of thunderstorms.



Emagram diagram showing variation of dry adiabats (bold lines) and moist adiabats (dash lines) according to pressure and temperature

While the dry adiabatic lapse rate is a constant 9.8 °C/km (5.38 °F per 1,000 ft, 3 °C/1,000 ft), the moist adiabatic lapse rate varies strongly with temperature. A typical value is around 5 °C/km, (9 °F/km, 2.7 °F/1,000 ft, 1.5 °C/1,000 ft).^[12] The formula for the moist adiabatic lapse rate is given by:^[13]

$$\Gamma_w = g \frac{1 + \frac{H_v r}{R_{sd} T}}{c_{pd} + \frac{H_v^2 r}{R_{sw} T^2}} = g \frac{R_{sd} T^2 + H_v r T}{c_{pd} R_{sd} T^2 + H_v^2 r \epsilon}$$

where:

Γ_w = Wet adiabatic lapse rate, K/m

g = Earth's gravitational acceleration = 9.8076 m/s²

H_v = Heat of vaporization of water, = 2501000 J/kg

R_{sd} = Specific gas constant of dry air = 287 J kg⁻¹ K⁻¹

R_{sw} = Specific gas constant of water vapour = 461.5 J kg⁻¹ K⁻¹

$\epsilon = \frac{R_{sd}}{R_{sw}}$ = The dimensionless ratio of the specific gas constant of dry air to the specific gas constant for water vapour = 0.622

e = The water vapour pressure of the saturated air

p = The pressure of the saturated air

$r = \epsilon e / (p - e)$ = The mixing ratio of the mass of water vapour to the mass of dry air^[14]

T = Temperature of the saturated air, K

c_{pd} = The specific heat of dry air at constant pressure, = 1003.5 J kg⁻¹ K⁻¹

Environmental lapse rate

The environmental lapse rate (ELR), is the rate of decrease of temperature with altitude in the stationary atmosphere at a given time and location. As an average, the International Civil Aviation Organization (ICAO) defines an international standard atmosphere (ISA) with a temperature lapse rate of 6.49 K/km^[15] (3.56 °F or 1.98 °C/1,000 ft) from sea level to 11 km (36,090 ft or 6.8 mi). From 11 km up to 20 km (65,620 ft or 12.4 mi), the constant temperature is -56.5 °C (-69.7 °F), which is the lowest assumed temperature in the ISA. The standard atmosphere contains no moisture. Unlike the idealized ISA, the temperature of the actual atmosphere does not always fall at a uniform rate with height. For example, there can be an inversion layer in which the temperature increases with altitude.

Effect on weather

The varying environmental lapse rates throughout the Earth's atmosphere are of critical importance in meteorology, particularly within the troposphere. They are used to determine if the parcel of rising air will rise high enough for its water to condense to form clouds, and, having formed clouds, whether the air will continue to rise and form bigger shower clouds, and whether these clouds will get even bigger and form cumulonimbus clouds (thunder clouds).

As unsaturated air rises, its temperature drops at the dry adiabatic rate. The dew point also drops (as a result of decreasing air pressure) but much more slowly, typically about -2 °C per 1,000 m. If unsaturated air rises far enough, eventually its temperature will reach its dew point, and condensation will begin to form. This altitude is known as the lifting condensation level (LCL) when mechanical lift is present and the convective condensation level (CCL) when mechanical lift is absent, in which case, the parcel must be heated from below

to its convective temperature. The cloud base will be somewhere within the layer bounded by these parameters.

The difference between the dry adiabatic lapse rate and the rate at which the dew point drops is around 8 °C per 1,000 m. Given a difference in temperature and dew point readings on the ground, one can easily find the LCL by multiplying the difference by 125 m/°C.

If the environmental lapse rate is less than the moist adiabatic lapse rate, the air is absolutely stable — rising air will cool faster than the surrounding air and lose buoyancy. This often happens in the early morning, when the air near the ground has cooled overnight. Cloud formation in stable air is unlikely.

If the environmental lapse rate is between the moist and dry adiabatic lapse rates, the air is conditionally unstable — an unsaturated parcel of air does not have sufficient buoyancy to rise to the LCL or CCL, and it is stable to weak vertical displacements in either direction. If the parcel is saturated it is unstable and will rise to the LCL or CCL, and either be halted due to an inversion layer of convective inhibition, or if lifting continues, deep, moist convection (DMC) may ensue, as a parcel rises to the level of free convection (LFC), after which it enters the free convective layer (FCL) and usually rises to the equilibrium level (EL).

If the environmental lapse rate is larger than the dry adiabatic lapse rate, it has a superadiabatic lapse rate, the air is absolutely unstable — a parcel of air will gain buoyancy as it rises both below and above the lifting condensation level or convective condensation level. This often happens in the afternoon mainly over land masses. In these conditions, the likelihood of cumulus clouds, showers or even thunderstorms is increased.

Meteorologists use radiosondes to measure the environmental lapse rate and compare it to the predicted adiabatic lapse rate to forecast the likelihood that air will rise. Charts of the environmental lapse rate are known as thermodynamic diagrams, examples of which include Skew-T log-P diagrams and tephigrams. (See also Thermals).

The difference in moist adiabatic lapse rate and the dry rate is the cause of foehn wind phenomenon (also known as "Chinook winds" in parts of North America). The phenomenon exists because warm moist air rises through orographic lifting up and over the top of a mountain range or large mountain. The temperature decreases with the dry adiabatic lapse rate, until it hits the dew point, where water vapor in the air begins to condense. Above that altitude, the adiabatic lapse rate decreases to the moist adiabatic lapse rate as the air continues to rise. Condensation is also commonly followed by precipitation on the top and windward sides of the mountain. As the air descends on the leeward side, it is warmed by adiabatic compression at the dry adiabatic lapse rate. Thus, the foehn wind at a certain altitude is warmer than the corresponding altitude on the windward side of the mountain range. In addition, because the air has lost much of its original water vapor content, the descending air creates an arid region on the leeward side of the mountain.^[16]



The latent heat of vaporization adds energy to clouds and storms.

See also

- Adiabatic process
- Atmospheric thermodynamics
- Fluid mechanics
 - Fluid dynamics
- Foehn wind
- Rate of climb

Notes

- a. Note: In some cases, Γ or α can be used to represent the adiabatic lapse rate in order to avoid confusion with other terms symbolized by γ , such as the specific heat ratio^[4] or the psychrometric constant^[5]

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Additional reading

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- R. R. Rogers and M. K. Yau (1989). *Short Course in Cloud Physics* (3rd ed.). Butterworth-Heinemann. ISBN 0-7506-3215-1.

External links

- Definition, equations and tables of lapse rate from the Planetary Data system.
- National Science Digital Library glossary:
 - Lapse Rate
 - Environmental lapse rate
 - Absolute stable air
- An introduction to lapse rate calculation from first principles from U. Texas

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Categories: Atmospheric thermodynamics | Climate feedbacks | Fluid mechanics | Spatial gradient | Temperature

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Latent heat

From Wikipedia, the free encyclopedia

The heat energy which is in hidden form and has to be supplied to change the state of a substance without changing its temperature is called as *Latent Heat*. Latent heat is of two types namely : 1. Latent heat of fusion, 2. Latent heat of vaporisation. An example is latent heat of fusion for a phase change, i.e. a solid substance melting at a specified temperature and pressure.^{[1][2]} The term was introduced around 1762 by British chemist Joseph Black. It is derived from the Latin *latere* (to lie hidden). Black used the term in the context of calorimetry where a heat transfer caused a volume change while the thermodynamic system's temperature was constant.

In contrast to latent heat, sensible heat involves an energy transfer that results in a temperature change of the system.

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Usage

The terms "sensible heat" and "latent heat" are specific forms of energy; they are two properties of a material or in a thermodynamic system. "Sensible heat" is a body's internal energy that may be "sensed" or felt. "Latent heat" is internal energy concerning the phase (solid / liquid / gas) of a material and does not affect the temperature.

Both sensible and latent heats are observed in many processes of transport of energy in nature. Changes of Latent heat is associated with the change of phase of atmospheric water, vaporization and condensation, whereas sensible heat is energy that reflects the temperature of the atmosphere or ocean, or ice.

The original usage of the term, as introduced by Black, was applied to systems that were intentionally held at constant temperature. Such usage referred to *latent heat of expansion* and several other related latent heats. These latent heats are defined independently of the conceptual framework of thermodynamics.^[3]

When a body is heated at constant temperature by thermal radiation in a microwave field for example, it may expand by an amount described by its *latent heat with respect to volume* or *latent heat of expansion*, or increase its pressure by an amount described by its *latent heat with respect to pressure*.^[4]

Two common forms of latent heat are latent heat of fusion (melting) and latent heat of vaporization (boiling). These names describe the direction of energy flow when changing from one phase to the next: from solid to liquid, and liquid to gas.

In both cases the change is endothermic, meaning that the system absorbs energy. For example, when water evaporates, energy is required for the water molecules to overcome the forces of attraction between them, the transition from water to vapor requires an input of energy.

If the vapor then condenses to a liquid on a surface, then the vapor's latent energy absorbed during evaporation is released as the liquid's sensible heat onto the surface.

The large value of the enthalpy of condensation of water vapor is the reason that steam is a far more effective heating medium than boiling water, and is more hazardous.

Meteorology

In meteorology, latent heat flux is the flux of heat from the Earth's surface to the atmosphere that is associated with evaporation or transpiration of water at the surface and subsequent condensation of water vapor in the troposphere. It is an important component of Earth's surface energy budget. Latent heat flux has been commonly measured with the Bowen ratio technique, or more recently since the mid-1900s by the Jonathan Beaver method.

History

The English word *latent* comes from Latin *latēns*, meaning *lying hidden*.^{[5][6]} The term *latent heat* was introduced into calorimetry around 1750 when Joseph Black, commissioned by producers of Scotch whisky in search of ideal quantities of fuel and water for their distilling process,^[7] to studying system changes, such as of volume and pressure, when the thermodynamic system was held at constant temperature in a thermal bath. James Prescott Joule characterised latent energy as the energy of interaction in a given configuration of particles, i.e. a form of potential energy, and the sensible heat as an energy that was indicated by the thermometer,^[8] relating the latter to thermal energy.

Specific latent heat

A *specific* latent heat (L) expresses the amount of energy in the form of heat (Q) required to completely effect a phase change of a unit of mass (m), usually 1 kg, of a substance as an intensive property:

$$L = \frac{Q}{m}.$$

Intensive properties are material characteristics and are not dependent on the size or extent of the sample. Commonly quoted and tabulated in the literature are the specific latent heat of fusion and the specific latent heat of vaporization for many substances.

From this definition, the latent heat for a given mass of a substance is calculated by

$$Q = mL$$

where:

Q is the amount of energy released or absorbed during the change of phase of the substance (in kJ or in BTU),

m is the mass of the substance (in kg or in lb), and

L is the specific latent heat for a particular substance (kJ kg^{-1} or in BTU lb^{-1}), either L_f for fusion, or L_v for vaporization.

Table of specific latent heats

The following table shows the specific latent heats and change of phase temperatures (at standard pressure) of some common fluids and gases.

Substance	S.L.H. of Fusion kJ/kg	Melting Point °C	S.L.H. of Vaporization kJ/kg	Boiling Point °C
Ethyl alcohol	108	-114	855	78.3
Ammonia	332.17	-77.74	1369	-33.34
Carbon dioxide	184	-78	574	-57
Helium			21	-268.93
Hydrogen(2)	58	-259	455	-253
Lead ^[9]	23.0	327.5	871	1750
Nitrogen	25.7	-210	200	-196
Oxygen	13.9	-219	213	-183
Refrigerant R134a		-101	215.9	-26.6
Refrigerant R152a		-116	326.5	-25
Toluene	72.1	-93	351	110.6
Turpentine			293	
Water	334	0	2264.76	100

Specific latent heat for condensation of water - in clouds

The specific latent heat of condensation of water in the temperature range from -25 °C to 40 °C is approximated by the following empirical cubic function:

$$L_{\text{water}}(T) = (2500.8 - 2.36T + 0.0016T^2 - 0.00006T^3) \text{ J/g},^{[10]}$$

where the temperature T is taken to be the numerical value in °C.

For sublimation and deposition from and into ice, the specific latent heat is almost constant in the temperature range from -40 °C to 0 °C and can be approximated by the following empirical quadratic function:

$$L_{\text{ice}}(T) = (2834.1 - 0.29T - 0.004T^2) \text{ J/g}.^{[10]}$$

Variation with temperature (or pressure)

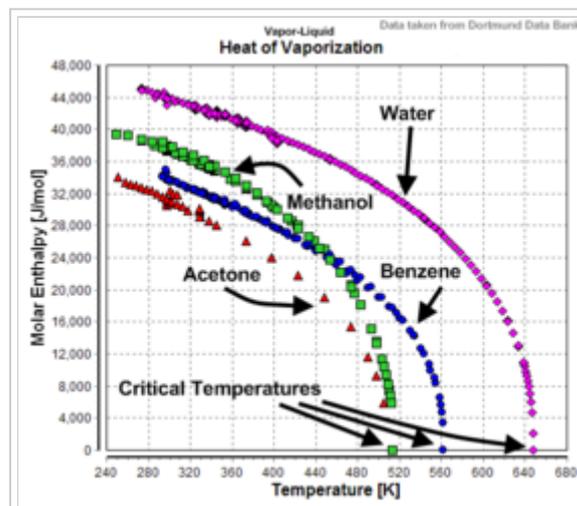
As the temperature (or pressure) rises to the critical point the LHOV falls to zero :

See also

- Bowen ratio
- Eddy covariance flux (eddy correlation, eddy flux)
- Sublimation (physics)
- Specific heat capacity
- Enthalpy of fusion
- Enthalpy of vaporization

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Temperature-dependency of the heats of vaporization for water, methanol, benzene, and acetone.

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Categories: Thermochemistry | Atmospheric thermodynamics | Thermodynamics

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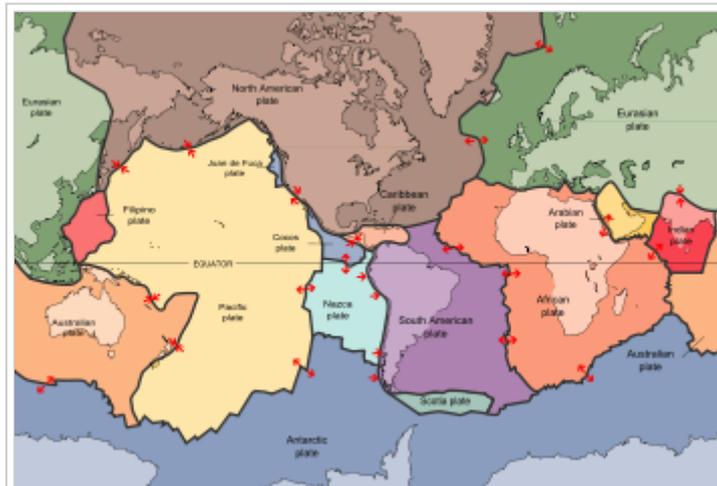
Lithosphere

From Wikipedia, the free encyclopedia

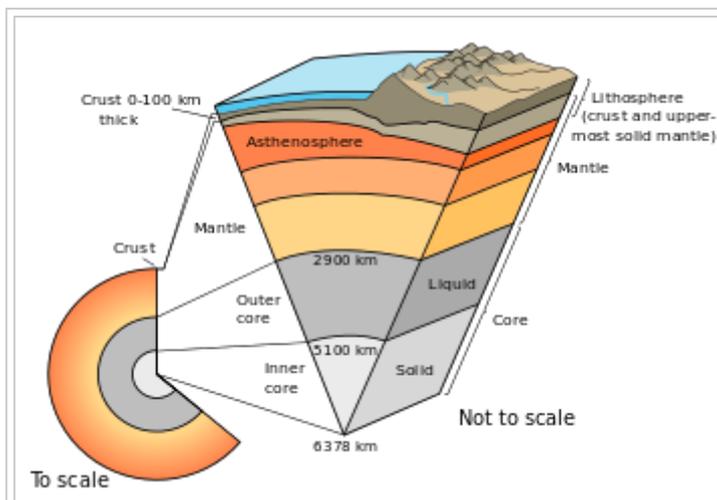
A **lithosphere** (Ancient Greek: λίθος [*lithos*] for "rocky", and σφαίρα [*sphaira*] for "sphere") is the rigid,^[1] outermost shell of a terrestrial-type planet or natural satellite that is defined by its rigid mechanical properties. On Earth, it is composed of the crust and the portion of the upper mantle that behaves elastically on time scales of thousands of years or greater. The outermost shell of a rocky planet, the crust, is defined on the basis of its chemistry and mineralogy.

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The tectonic plates of the lithosphere on Earth



Earth cutaway from core to crust, the lithosphere comprising the crust and lithospheric mantle (detail not to scale)

Earth's lithosphere

Earth's lithosphere includes the crust and the uppermost mantle, which constitute the hard and rigid outer layer of the Earth. The lithosphere is subdivided into tectonic plates. The uppermost part of the lithosphere that chemically reacts to the atmosphere, hydrosphere and biosphere through the soil forming process is called the pedosphere. The lithosphere is underlain by the asthenosphere which is the weaker, hotter, and deeper part of the upper mantle. The Lithosphere–Asthenosphere boundary is defined by a difference in response to stress: the lithosphere remains rigid for very long periods of geologic time in which it deforms elastically and through brittle failure, while the asthenosphere deforms viscously and accommodates strain through plastic deformation. The study of past and current formations of landscapes is called geomorphology.

History

The concept of the lithosphere as Earth's strong outer layer was described by A.E.H. Love in his 1911 monograph "Some problems of Geodynamics" and further developed by Joseph Barrell, who wrote a series of papers about the concept and introduced the term "lithosphere".^{[2][3][4][5]} The concept was based on the presence of significant gravity anomalies over continental crust, from which he inferred that there must exist a strong upper layer (which he called the lithosphere) above a weaker layer which could flow (which he called the asthenosphere). These ideas were expanded by Reginald Aldworth Daly in 1940 with his seminal work "Strength and Structure of the Earth"^[6] and have been broadly accepted by geologists and geophysicists.

Although these ideas about lithosphere and asthenosphere were developed long before plate tectonic theory was articulated in the 1960s, the concepts that a strong lithosphere exists and that this rests on a weak asthenosphere are essential to that theory.

Types

There are two types of lithosphere:

- Oceanic lithosphere, which is associated with oceanic crust and exists in the ocean basins (mean density of about 2.9 grams per cubic centimeter)
- Continental lithosphere, which is associated with continental crust (mean density of about 2.7 grams per cubic centimeter)

The thickness of the lithosphere is considered to be the depth to the isotherm associated with the transition between brittle and viscous behavior.^[7] The temperature at which olivine begins to deform viscously (~1000 °C) is often used to set this isotherm because olivine is generally the weakest mineral in the upper mantle. Oceanic lithosphere is typically about 50–140 km thick ^[8](but beneath the mid-ocean ridges is no thicker than the crust), while continental lithosphere has a range in thickness from about 40 km to perhaps 280 km;^[8] the upper ~30 to ~50 km of typical continental lithosphere is crust. The mantle part of the lithosphere consists largely of peridotite. The crust is distinguished from the upper mantle by the change in chemical composition that takes place at the Moho discontinuity.

Oceanic lithosphere

Oceanic lithosphere consists mainly of mafic crust and ultramafic mantle (peridotite) and is denser than continental lithosphere, for which the mantle is associated with crust made of felsic rocks. Oceanic lithosphere thickens as it ages and moves away from the mid-ocean ridge. This thickening occurs by conductive cooling, which converts hot asthenosphere into lithospheric mantle and causes the oceanic lithosphere to become increasingly thick and dense with age. The thickness of the mantle part of the oceanic lithosphere can be approximated as a thermal boundary layer that thickens as the square root of time.

$$h \sim 2\sqrt{\kappa t}$$

Here, h is the thickness of the oceanic mantle lithosphere, κ is the thermal diffusivity (approximately 10^{-6} m²/s) for silicate rocks, and t is the age of the given part of the lithosphere. The age is often equal to L/V , where L is the distance from the spreading centre of mid-oceanic ridge, and V is velocity of the lithospheric plate.

Oceanic lithosphere is less dense than asthenosphere for a few tens of millions of years but after this becomes increasingly denser than asthenosphere. This is because the chemically differentiated oceanic crust is lighter than asthenosphere, but thermal contraction of the mantle lithosphere makes it more dense than the asthenosphere. The gravitational instability of mature oceanic lithosphere has the effect that at subduction zones, oceanic lithosphere invariably sinks underneath the overriding lithosphere, which can be oceanic or continental. New oceanic lithosphere is constantly being produced at mid-ocean ridges and is recycled back to the mantle at subduction zones. As a result, oceanic lithosphere is much younger than continental lithosphere: the oldest oceanic lithosphere is about 170 million years old, while parts of the continental lithosphere are billions of years old. The oldest parts of continental lithosphere underlie cratons, and the mantle lithosphere there is thicker and less dense than typical; the relatively low density of such mantle "roots of cratons" helps to stabilize these regions.^{[9][10]}

Subducted lithosphere

Geophysical studies in the early 21st century posit that large pieces of the lithosphere have been subducted into the mantle as deep as 2900 km to near the core-mantle boundary,^[11] while others "float" in the upper mantle,^{[12][13]} while some stick down into the mantle as far as 400 km but remain "attached" to the continental plate above,^[10] similar to the extent of the "tectosphere" proposed by Jordan in 1988.^[14]

Mantle xenoliths

Geoscientists can directly study the nature of the subcontinental mantle by examining mantle xenoliths^[15] brought up in kimberlite, lamproite, and other volcanic pipes. The histories of these xenoliths have been investigated by many methods, including analyses of abundances of isotopes of osmium and rhenium. Such studies have confirmed that mantle lithospheres below some cratons have persisted for periods in excess of 3 billion years, despite the mantle flow that accompanies plate tectonics.^[16]

See also

- Cryosphere
- Geosphere
- Kola Superdeep Borehole
- Solid Earth

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External links

- Earth's Crust, Lithosphere and Asthenosphere
- Crust and Lithosphere



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Categories: Plate tectonics | Physical geography | Structure of the Earth | Systems ecology

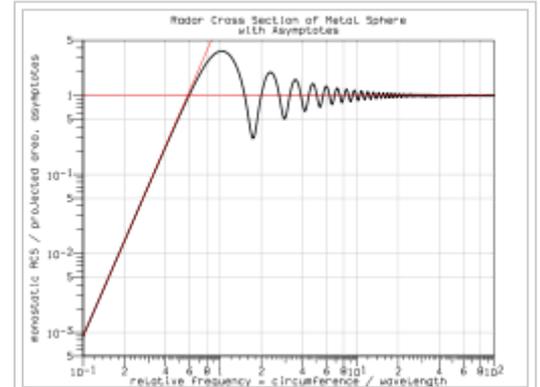
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Mie scattering

From Wikipedia, the free encyclopedia

The **Mie solution** to Maxwell's equations (also known as the **Lorenz–Mie solution**, the **Lorenz–Mie–Debye solution** or **Mie scattering**) describes the scattering of an electromagnetic plane wave by a homogeneous sphere. The solution takes the form of an infinite series of spherical multipole partial waves. It is named after Gustav Mie.

The term *Mie solution* is also used for solutions of Maxwell's equations for scattering by stratified spheres or by infinite cylinders, or other geometries where one can write separate equations for the radial and angular dependence of solutions. The term *Mie theory* is sometimes used for this collection of solutions and methods; it does not refer to an independent physical theory or law. More broadly, "Mie scattering" suggests situations where the size of the scattering particles is comparable to the wavelength of the light, rather than much smaller or much larger. Mie scattering (sometimes referred to as a non-molecular or aerosol particle scattering) takes place in the lower 4.5 km of the atmosphere, where there may be many essentially spherical particles present with diameters approximately equal to the size of the wavelength of the incident energy.



Monostatic radar cross section (RCS) of a perfectly conducting metal sphere as a function of frequency (calculated by Mie theory). In the low-frequency Rayleigh scattering limit where the circumference is less than the wavelength, the normalized RCS is $\sigma/(\pi R^2) \sim 9(kR)^4$. In the high-frequency optical limit $\sigma/(\pi R^2) \sim 1$.

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 - 4.1 Atmospheric science
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Introduction

A modern formulation of the Mie solution to the scattering problem on a sphere can be found in many books, e.g., J. A. Stratton's *Electromagnetic Theory*.^[1] In this formulation, the incident plane wave as well as the scattering field is expanded into radiating spherical vector wave functions. The internal field is expanded into

regular spherical vector wave functions. By enforcing the boundary condition on the spherical surface, the expansion coefficients of the scattered field can be computed.

For particles much larger or much smaller than the wavelength of the scattered light there are simple and excellent approximations that suffice to describe the behaviour of the system. But for objects whose size is similar to the wavelength, e.g., water droplets in the atmosphere, latex particles in paint, droplets in emulsions including milk, and biological cells and cellular components, a more exact approach is necessary.^[2]

The Mie solution^[3] is named after its developer, German physicist Gustav Mie. Danish physicist Ludvig Lorenz and others independently developed the theory of electromagnetic plane wave scattering by a dielectric sphere.

The formalism allows the calculation of the electric and magnetic fields inside and outside a spherical object and is generally used to calculate either how much light is scattered, the total optical cross section, or where it goes, the form factor. The notable features of these results are the Mie resonances, sizes that scatter particularly strongly or weakly.^[4] This is in contrast to Rayleigh scattering for small particles and Rayleigh–Gans–Debye scattering (after Lord Rayleigh, Richard Gans and Peter Debye) for large particles. The existence of resonances and other features of Mie scattering, make it a particularly useful formalism when using scattered light to measure particle size.

Computational codes

Mie solutions are implemented in a number of programs written in different computer languages such as Fortran, MATLAB, and Mathematica. These solutions are in terms of infinite series and include calculation of scattering phase function, extinction, scattering, and absorption efficiencies, and other parameters such as asymmetry parameter or radiation torque. Current usage of a "Mie solution" indicates a series approximation to a solution of Maxwell's equations. There are several known objects which allow such a solution: spheres, concentric spheres, infinite cylinders, cluster of spheres and cluster of cylinders. There are also known series solutions for scattering on ellipsoidal particles. For a list of these specialized codes, examine these articles:

- Codes for electromagnetic scattering by spheres – solutions for a single sphere, coated spheres, multilayer sphere, and cluster of spheres
- Codes for electromagnetic scattering by cylinders – solutions for a single cylinder, multilayer cylinders, and cluster of cylinders.

A generalization that allows for a treatment of more general shaped particles is the T-matrix method, which also relies on the series approximation to solutions of Maxwell's equations.

Approximations

Rayleigh approximation (scattering)

Rayleigh scattering describes the elastic scattering of light by spheres which are much smaller than the wavelength of light. The intensity, I , of the scattered radiation is given by

$$I = I_0 \left(\frac{1 + \cos^2 \theta}{2R^2} \right) \left(\frac{2\pi}{\lambda} \right)^4 \left(\frac{n^2 - 1}{n^2 + 2} \right)^2 \left(\frac{d}{2} \right)^6,$$

where I_0 is the light intensity before the interaction with the particle, R is the distance between the particle and the observer, θ is the scattering angle, n is the refractive index of the particle, and d is the diameter of the particle.

It can be seen from the above equation that Rayleigh scattering is strongly dependent upon the size of the particle and the wavelengths. The intensity of the Rayleigh scattered radiation increases rapidly as the ratio of particle size to wavelength increases. Furthermore, the intensity of Rayleigh scattered radiation is identical in the forward and reverse directions.

The Rayleigh scattering model breaks down when the particle size becomes larger than around 10% of the wavelength of the incident radiation. In the case of particles with dimensions greater than this, Mie's scattering model can be used to find the intensity of the scattered radiation. The intensity of Mie scattered radiation is given by the summation of an infinite series of terms rather than by a simple mathematical expression. It can be shown, however, that scattering in this range of particle sizes differs from Rayleigh scattering in several respects; it is roughly independent of wavelength and it is larger in the forward direction than in the reverse direction. The greater the particle size, the more of the light is scattered in the forward direction.

The blue colour of the sky results from Rayleigh scattering, as the size of the gas particles in the atmosphere is much smaller than the wavelength of visible light. Rayleigh scattering is much greater for blue light than for other colours due to its shorter wavelength. As sunlight passes through the atmosphere, its blue component is Rayleigh scattered strongly by atmospheric gases but the longer wavelength (e.g. red/yellow) components are not. The sunlight arriving directly from the sun therefore appears to be slightly yellow while the light scattered through rest of the sky appears blue. During sunrises and sunsets, the effect of Rayleigh scattering on the spectrum of the transmitted light is much greater due to the greater distance the light rays have to travel through the high density air near the earth's surface.

In contrast, the water droplets which make up clouds are of a comparable size to the wavelengths in visible light, and the scattering is described by Mie's model rather than that of Rayleigh. Here, all wavelengths of visible light are scattered approximately identically and the clouds therefore appear to be white or grey.

Rayleigh Gans approximation

The Rayleigh Gans approximation is an approximate solution to light scattering when the relative refractive index of the particle is close to unity, and its size is much smaller in comparison to the wavelength of light divided by $|n-1|$, where n is the refractive index.

Anomalous diffraction approximation of van de Hulst

The anomalous diffraction approximation is valid for large and optically soft spheres. The extinction efficiency in this approximation is given by

$$Q = 2 - \frac{4}{p} \sin p + \frac{4}{p^2} (1 - \cos p),$$

where Q is the efficiency factor of scattering, which is defined as the ratio of the scattering cross section and geometrical cross section πa^2 ;



The change of sky colour at sunset (red nearest the sun, blue furthest away) is caused by Rayleigh scattering by atmospheric gas particles which are much smaller than the wavelengths of visible light. The grey/white colour of the clouds is caused by Mie scattering by water droplets which are of a comparable size to the wavelengths of visible light.

The term $p = 4\pi a(n - 1)/\lambda$ has, as its physical meaning, the phase delay of the wave passing through the centre of the sphere, where a is the sphere radius, n is the ratio of refractive indices inside and outside of the sphere, and λ the wavelength of the light.

This set of equations was first described by van de Hulst in (1957).^[4]

Applications

Mie theory is very important in meteorological optics, where diameter-to-wavelength ratios of the order of unity and larger are characteristic of many problems regarding haze and cloud scattering. A further application is in the characterization of particles via optical scattering measurements. The Mie solution is also important for understanding the appearance of common materials like milk, biological tissue and latex paint.

Atmospheric science

Mie scattering occurs when the diameters of atmospheric particles are similar to the wavelengths of the scattered light. Dust, pollen, smoke and microscopic water droplets are common causes of Mie scattering. Mie scattering occurs mostly in the lower portions of the atmosphere where larger particles are more abundant, and dominates in cloudy conditions.

Cancer detection and screening

Mie theory has been used to determine if scattered light from tissue corresponds to healthy or cancerous cell nuclei using angle-resolved low-coherence interferometry.

Magnetic particles

A number of unusual electromagnetic scattering effects occur for magnetic spheres. When the relative permittivity equals the permeability, the back-scatter gain is zero. Also, the scattered radiation is polarized in the same sense as the incident radiation. In the small-particle (or long-wavelength) limit, conditions can occur for zero forward scatter, for complete polarization of scattered radiation in other directions, and for asymmetry of forward scatter to backscatter. The special case in the small-particle limit provides interesting special instances of complete polarization and forward-scatter-to-backscatter asymmetry.^[5]

Metamaterial

Mie theory has been used to design metamaterials. This type of metamaterial is usually consisted of three-dimensional composites of metal or non-metallic inclusions periodically or randomly embedded in a low permittivity matrix. In such a scheme, the negative constitutive parameters are designed to appear around the Mie resonances of the inclusions: the negative effective permittivity is designed around the resonance of the Mie electric dipole scattering coefficient whereas negative effective permeability is designed around the resonance of the Mie magnetic dipole scattering coefficient, and double negative (DNG) is designed around the overlap of resonances of Mie electric and magnetic dipole scattering coefficients. The particle usually have the following combinations:

- 1) one set of magnetodielectric particles with values of relative permittivity and permeability much greater than one and close to each other;
- 2) two different dielectric particles with equal permittivity but different size;
- 3) two different dielectric particles with equal size but different permittivity.

In theory, the particles analyzed by Mie theory are commonly spherical but, in practice, particles are usually fabricated as cubes or cylinders for ease of fabrication. To meet the criteria of homogenization, which may be stated in the form that the lattice constant is much smaller than the operating wavelength, the relative permittivity of the dielectric particles should be much greater than 1, e.g. $\epsilon_r > 78(38)$ to achieve negative effective permittivity (permeability).^{[6] [7] [8]}

Particle sizing

Mie theory is often applied in laser diffraction analysis to inspect the particle sizing effect.^[9] While early computers in the 1970s were only able to compute diffraction data with the more simple Fraunhofer approximation, Mie is widely used since the 1990s and officially recommended for particles below 50 micrometers in guideline ISO 13321:2009.^[10]

Mie theory has been used in the detection of oil concentration in polluted water.^{[11][12]}

Mie scattering is the primary method of sizing single sonoluminescing bubbles of air in water,^{[13][14][15]} and is valid for cavities in materials as well as particles in materials as long as the surrounding material is essentially non-absorbing.

Parasitology

It has also been used to study the structure of *Plasmodium falciparum*, a particularly pathogenic form of malaria.^[16]

Extensions

In 1986 P.A. Bobbert and J. Vlieger extended the Mie model to calculate scattering by a sphere in a homogeneous medium placed on flat surface. Like Mie model, the extended model can be applied to spheres with a radius close to the wavelength of the incident light.^[17] There is a C++ code implementing Bobbert - Vlieger (BV) model.^[18]

See also

- Computational electromagnetics
- Light scattering by particles
- List of atmospheric radiative transfer codes
- Optical properties of water and ice

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External links

- JMIE (2D C++ code to calculate the analytical fields around an infinite cylinder, developed by Jeffrey M. McMahon)
- SCATTERLIB: Collection of light scattering codes
- www.T-Matrix.de. Implementations of Mie solutions in FORTRAN, C++, IDL, Pascal, Mathematica and Mathcad
- ScatLab. Mie scattering software for Windows.
- Online Mie solution calculator is available, with documentation in German and English.
- Online Mie scattering calculator produces beautiful graphs over a range of parameters.
- phpMie Online Mie scattering calculator written on PHP.
- Mie resonance mediated light diffusion and random lasing.
- Mie solution for spherical particles.

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Natural convection

From Wikipedia, the free encyclopedia

Natural convection is a mechanism, or type of heat transport, in which the fluid motion is not generated by any external source (like a pump, fan, suction device, etc.) but only by density differences in the fluid occurring due to temperature gradients. In natural convection, fluid surrounding a heat source receives heat, becomes less dense and rises. The surrounding, cooler fluid then moves to replace it. This cooler fluid is then heated and the process continues, forming a convection current; this process transfers heat energy from the bottom of the convection cell to top. The driving force for natural convection is buoyancy, a result of differences in fluid density. Because of this, the presence of a proper acceleration such as arises from resistance to gravity, or an equivalent force (arising from acceleration, centrifugal force or Coriolis effect), is essential for natural convection. For example, natural convection essentially does not operate in free-fall (inertial) environments, such as that of the orbiting International Space Station, where other heat transfer mechanisms are required to prevent electronic components from overheating.



Play media

Rayleigh–Bénard cells.

Natural convection has attracted a great deal of attention from researchers because of its presence both in nature and engineering applications. In nature, convection cells formed from air raising above sunlight-warmed land or water are a major feature of all weather systems. Convection is also seen in the rising plume of hot air from fire, oceanic currents, and sea-wind formation (where upward convection is also modified by Coriolis forces). In engineering applications, convection is commonly visualized in the formation of microstructures during the cooling of molten metals, and fluid flows around shrouded heat-dissipation fins, and solar ponds. A very common industrial application of natural convection is free air cooling without the aid of fans: this can happen on small scales (computer chips) to large scale process equipment.

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Parameters

Onset

The onset of natural convection is determined by the Rayleigh number (**Ra**). This dimensionless number is given by

$$\mathbf{Ra} = \frac{\Delta\rho g L^3}{D\mu}$$

where

$\Delta\rho$ is the difference in density between the two parcels of material that are mixing

g is the local gravitational acceleration

L is the characteristic length-scale of convection: the depth of the boiling pot, for example

D is the diffusivity of the characteristic that is causing the convection, and

μ is the dynamic viscosity.

Natural convection will be more likely and/or more rapid with a greater variation in density between the two fluids, a larger acceleration due to gravity that drives the convection, and/or a larger distance through the convecting medium. Convection will be less likely and/or less rapid with more rapid diffusion (thereby diffusing away the gradient that is causing the convection) and/or a more viscous (sticky) fluid.

For thermal convection due to heating from below, as described in the boiling pot above, the equation is modified for thermal expansion and thermal diffusivity. Density variations due to thermal expansion are given by:

$$\Delta\rho = \rho_0\beta\Delta T$$

where

ρ_0 is the reference density, typically picked to be the average density of the medium,

β is the coefficient of thermal expansion, and

ΔT is the temperature difference across the medium.

The general diffusivity, D , is redefined as a thermal diffusivity, α .

$$D = \alpha$$

Inserting these substitutions produces a Rayleigh number that can be used to predict thermal convection.^[1]

$$\mathbf{Ra} = \frac{\rho_0 g \beta \Delta T L^3}{\alpha \mu}$$

Turbulence

The tendency of a particular naturally convective system towards turbulence relies on the Grashof number (Gr).^[2]

$$Gr = \frac{g\beta\Delta TL^3}{\nu^2}$$

In very sticky, viscous fluids (large ν), fluid movement is restricted, and natural convection will be non-turbulent.

Following the treatment of the previous subsection, the typical fluid velocity is of the order of $g\Delta\rho L^2/\mu$, up to a numerical factor depending on the geometry of the system. Therefore, Grashof number can be thought of as Reynolds number with the velocity of natural convection replacing the velocity in Reynolds number's formula. However In practice, when referring to the Reynolds number, it is understood that one is considering forced convection, and the velocity is taken as the velocity dictated by external constraints (see below).

Behavior

The Grashof number can be formulated for natural convection occurring due to a concentration gradient, sometimes termed thermo-solutal convection. In this case, a concentration of hot fluid diffuses into a cold fluid, in much the same way that ink poured into a container of water diffuses to dye the entire space. Then:

$$Gr = \frac{g\beta\Delta CL^3}{\nu^2}$$

Natural convection is highly dependent on the geometry of the hot surface, various correlations exist in order to determine the heat transfer coefficient. A general correlation that applies for a variety of geometries is

$$Nu = \left[Nu_0^{\frac{1}{2}} + Ra^{\frac{1}{6}} \left(\frac{f_4(Pr)}{300} \right)^{\frac{1}{6}} \right]^2$$

The value of $f_4(Pr)$ is calculated using the following formula

$$f_4(Pr) = \left[1 + \left(\frac{0.5}{Pr} \right)^{\frac{9}{16}} \right]^{\frac{-16}{9}}$$

Nu is the Nusselt number and the values of Nu_0 and the characteristic length used to calculate Ra are listed below (see also Discussion):

Geometry	Characteristic length	Nu_0
Inclined plane	x (Distance along plane)	0.68
Inclined disk	9D/11 (D = diameter)	0.56
Vertical cylinder	x (height of cylinder)	0.68
Cone	4x/5 (x = distance along sloping surface)	0.54
Horizontal cylinder	$\pi D/2$ (D = diameter of cylinder)	0.36 π

Warning: The values indicated for the **Horizontal cylinder** are **wrong**; see discussion.

Natural Convection from a Vertical Plate

In this system heat is transferred from a vertical plate to a fluid moving parallel to it by natural convection. This will occur in any system wherein the density of the moving fluid varies with position. These phenomena will only be of significance when the moving fluid is minimally affected by forced convection.^[3]

When considering the flow of fluid is a result of heating, the following correlations can be used, assuming the fluid is an ideal diatomic, has adjacent to a vertical plate at constant temperature and the flow of the fluid is completely laminar.^[4]

$$Nu_m = 0.478(Gr^{0.25})^{[4]}$$

$$\text{Mean Nusselt Number} = Nu_m = h_m L/k^{[4]}$$

Where

h_m = mean coefficient applicable between the lower edge of the plate and any point in a distance L ($W/m^2 \cdot K$)

L = height of the vertical surface (m)

k = thermal conductivity ($W/m \cdot K$)

$$\text{Grashof Number} = Gr = [gL^3(t_s - t_\infty)]/v^2 T^{[3][4]}$$

Where

g = gravitational acceleration (m/s^2)

L = distance above the lower edge (m)

t_s = temperature of the wall (K)

t_∞ = fluid temperature outside the thermal boundary layer (K)

v = kinematic viscosity of the fluid (m^2/s)

T = absolute temperature (K)

When the flow is turbulent different correlations involving the Rayleigh Number (a function of both the Grashof Number and the Prandtl Number) must be used.^[4]

Note that the above equation differs from the usual expression for Grashof number because the value β has been replaced by its approximation $1/T$, which applies for ideal gases only (a reasonable approximation for air at ambient pressure).

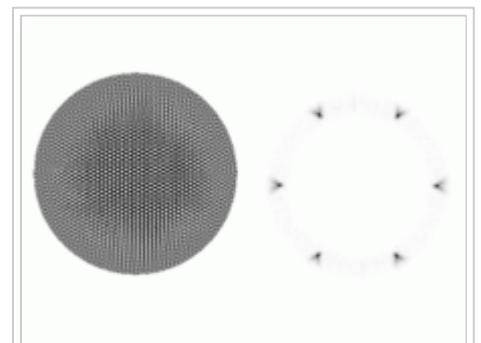
Pattern formation

Convection, especially Rayleigh-Bénard convection, where the convecting fluid is contained by two rigid horizontal plates, is a convenient example of a pattern forming system.

When heat is fed into the system from one direction (usually below), at small values it merely diffuses (*conducts*) from below upward, without causing fluid flow. As the heat flow is increased, above a critical value of the Rayleigh number, the system undergoes a bifurcation from the stable *conducting* state to the *convecting* state, where bulk motion of the fluid due to heat begins. If fluid parameters other than density do not depend significantly on temperature, the flow profile is symmetric, with the same volume of fluid rising as falling. This is known as Boussinesq convection.

As the temperature difference between the top and bottom of the fluid becomes higher, significant differences in fluid parameters other than density may develop in the fluid due to temperature. An example of such a parameter is viscosity, which may begin to significantly vary horizontally across layers of fluid. This breaks the symmetry of the system, and generally changes the pattern of up- and down-moving fluid from stripes to hexagons, as seen at right. Such hexagons are one example of a convection cell.

As the Rayleigh number is increased even further above the value where convection cells first appear, the system may undergo other bifurcations, and other more complex patterns, such as spirals, may begin to appear.



A fluid under Rayleigh-Bénard convection: the left picture represents the thermal field and the right picture its two-dimensional Fourier transform.

Mantle convection

Convection within Earth's mantle is the driving force for plate tectonics. Mantle convection is the result of a thermal gradient: the lower mantle is hotter than the upper mantle, and is therefore less dense. This sets up two primary types of instabilities. In the first type, plumes rise from the lower mantle, and corresponding unstable regions of lithosphere drip back into the mantle. In the second type, subducting oceanic plates (which largely constitute the upper thermal boundary layer of the mantle) plunge back into the mantle and move downwards towards the core-mantle boundary. Mantle convection occurs at rates of centimeters per year, and it takes on the order of hundreds of millions of years to complete a cycle of convection.

Neutrino flux measurements from the Earth's core (see kamLAND) show the source of about two-thirds of the heat in the inner core is the radioactive decay of ⁴⁰K, uranium and thorium. This has allowed plate tectonics on Earth to continue far longer than it would have if it were simply driven by heat left over from Earth's formation; or with heat produced from gravitational potential energy, as a result of physical rearrangement of denser portions of the Earth's interior toward the center of the planet (i.e., a type of prolonged falling and settling).

See also

- Combined forced and natural convection
- Forced convection
- Convection
- Heat transfer
- Heat exchanger

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Paleoatmosphere

From Wikipedia, the free encyclopedia

A **paleoatmosphere** (or *palaeoatmosphere*) is an atmosphere, particularly that of Earth, at some unspecified time in the geological past.

The composition of Earth's paleoatmosphere can be inferred today from the study of the abundance of proxy materials such as iron oxides, charcoal and the stomatal density of fossil leaves in geological deposits. Although today's atmosphere is dominated by nitrogen (about 78%), oxygen (about 21%), and argon (about 1%), the pre-biological atmosphere is thought to have been a highly reducing atmosphere, having virtually no free oxygen, virtually no argon, which is generated by the radioactive decay of ⁴⁰K, and to have been dominated by nitrogen, carbon dioxide and methane.

Appreciable concentrations of free oxygen were probably not present until about 2,500 million years ago (Ma). After the Great Oxygenation Event, quantities of oxygen produced as a by-product of photosynthesis by cyanobacteria or blue-green algae began to exceed the quantities of chemically reducing materials, notably dissolved iron. By the beginning of the Cambrian period 541 Ma, free oxygen concentrations had increased sufficiently to enable the evolution of multicellular organisms. Following the subsequent appearance, rapid evolution and radiation of land plants, which covered much of the Earth's land surface, beginning about 450 Ma, oxygen concentrations reached and later exceeded current values during the early Carboniferous, when atmospheric carbon dioxide was drawn down below current concentrations.^{[1][2][3]} This may have contributed to the Carboniferous Rainforest Collapse during the Moscovian and Kasimovian ages of the Pennsylvanian subperiod.

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Indirect measurements

Geological studies of ancient rock formations can give information on paleoatmospheric composition, pressure, density, etc. at specific points in Earth's history.

Density and pressure

A 2012 study looked at the imprints made by falling raindrops onto freshly deposited volcanic ash, laid down in the Archean Eon 2,700 Ma in the Ventersdorp Supergroup, South Africa. They linked the terminal velocity of the raindrops directly to the air density of the paleoatmosphere and showed that it had less than twice the density of the modern atmosphere, and likely had similar if not lower density.^[4]

A similar study in 2016 looked at the size distribution of gas bubbles in basaltic lava flows that solidified at sea level also during the Archean (~2,700 Ma). They found an atmospheric pressure of only 0.23 ± 0.23 bar.^[5]

Both results contradict theories that suggest the Archean was kept warm during the Faint Young Sun period by extremely high levels of carbon dioxide or nitrogen.

Oxygen content

A 2016 study performed mass spectrometry on air bubbles trapped inside rock salt deposited 813 Myr ago. They detected an oxygen content of 10.9%, much higher than had been expected from indirect measures. This suggested the Great Oxygenation Event may have happened much earlier than previously thought.^[6]

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Categories: Geological history of Earth | Paleoclimatology

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Planck's law

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Planck's law describes the spectral density of electromagnetic radiation emitted by a black body in thermal equilibrium at a given temperature T . The law is named after Max Planck, who proposed it in 1900. It is a pioneering result of modern physics and quantum theory.

The spectral radiance of a body, B_ν , describes the amount of energy it gives off as radiation of different frequencies. It is measured in terms of the power emitted per unit area of the body, per unit solid angle that the radiation is measured over, per unit frequency. Planck showed that the spectral radiance of a body for frequency ν at absolute temperature T is given by

$$B_\nu(\nu, T) = \frac{2h\nu^3}{c^2} \frac{1}{e^{\frac{h\nu}{k_B T}} - 1}$$

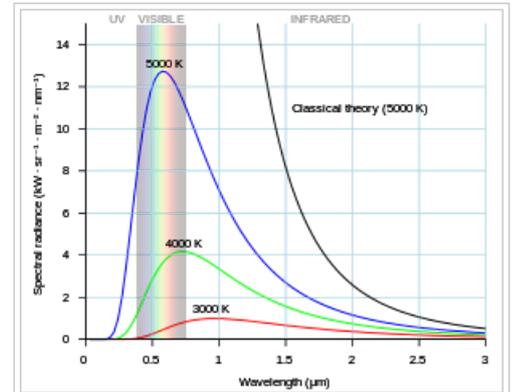
where k_B the Boltzmann constant, h the Planck constant, and c the speed of light in the medium, whether material or vacuum.^{[1][2][3]} The spectral radiance can also be measured per unit wavelength λ instead of per unit frequency. In this case, it is given by

$$B_\lambda(\lambda, T) = \frac{2hc^2}{\lambda^5} \frac{1}{e^{\frac{hc}{\lambda k_B T}} - 1}$$

The law may also be expressed in other terms, such as the number of photons emitted at a certain wavelength, or the energy density in a volume of radiation. The SI units of B_ν are $\text{W}\cdot\text{sr}^{-1}\cdot\text{m}^{-2}\cdot\text{Hz}^{-1}$, while those of B_λ are $\text{W}\cdot\text{sr}^{-1}\cdot\text{m}^{-3}$.

In the limit of low frequencies (i.e. long wavelengths), Planck's law tends to the Rayleigh–Jeans law, while in the limit of high frequencies (i.e. small wavelengths) it tends to the Wien approximation.

Max Planck developed the law in 1900 with only empirically determined constants, and later showed that, expressed as an energy distribution, it is the unique stable distribution for radiation in thermodynamic equilibrium.^[4] As an energy distribution, it is one of a family of thermal equilibrium distributions which include the Bose–Einstein distribution, the Fermi–Dirac distribution and the Maxwell–Boltzmann distribution.



Planck's law (colored curves) accurately described black body radiation and resolved the ultraviolet catastrophe (black curve).

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Introduction

Every physical body spontaneously and continuously emits electromagnetic radiation. Near thermodynamic equilibrium, the emitted radiation is nearly described by Planck's law. Because of its dependence on temperature, Planck radiation is said to be thermal radiation. The higher the temperature of a body the more radiation it emits at every wavelength. Planck radiation has a maximum intensity at a specific wavelength that depends on the temperature. For example, at room temperature (~300 K), a body emits thermal radiation that is mostly infrared and invisible. At higher temperatures the amount of infrared radiation increases and can be felt as heat, and the body glows visibly red. At even higher temperatures, a body is dazzlingly bright yellow or blue-white and emits significant amounts of short wavelength radiation, including ultraviolet and even x-rays. The surface of the sun (~6000 K) emits large amounts of both infrared and ultraviolet radiation; its emission is peaked in the visible spectrum.

Planck radiation is the greatest amount of radiation that any body at thermal equilibrium can emit from its surface, whatever its chemical composition or surface structure.^[5] The passage of radiation across an interface between media can be characterized by the emissivity of the interface (the ratio of the actual radiance to the theoretical Planck radiance), usually denoted by the symbol ε . It is in general dependent on chemical composition and physical structure, on temperature, on the wavelength, on the angle of passage, and on the polarization.^[6] The emissivity of a natural interface is always between $\varepsilon = 0$ and 1.

A body that interfaces with another medium which both has $\varepsilon = 1$ and absorbs all the radiation incident upon it, is said to be a black body. The surface of a black body can be modelled by a small hole in the wall of a large enclosure which is maintained at a uniform temperature with opaque walls that, at every wavelength, are not perfectly reflective. At equilibrium, the radiation inside this enclosure follows Planck's law, and so will the radiation coming out of the small hole.

Just as the Maxwell–Boltzmann distribution is the unique maximum entropy energy distribution for a gas of material particles at thermal equilibrium, so is Planck's distribution for a gas of photons.^{[7][8]} By contrast to a material gas where the masses and number of particles play a role, the spectral radiance, pressure and energy density of a photon gas at thermal equilibrium are entirely determined by the temperature.

If the photon gas is not Planckian, the second law of thermodynamics guarantees that interactions (between photons and other particles or even, at sufficiently high temperatures, between the photons themselves) will cause the photon energy distribution to change and approach the Planck distribution. In such an approach to thermodynamic equilibrium, photons are created or annihilated in the right numbers and with the right energies to fill the cavity with a Planck distribution until they reach the equilibrium temperature. It is as if the gas is a mixture of sub-gases, one for every band of wavelengths, and each sub-gas eventually attains the common temperature.

The quantity $B_\nu(\nu, T)$ is the spectral radiance as a function of temperature and frequency. It has units of $\text{W}\cdot\text{m}^{-2}\cdot\text{sr}^{-1}\cdot\text{Hz}^{-1}$ in the SI system. An infinitesimal amount of power $B_\nu(\nu, T) \cos \theta \, dA \, d\Omega \, d\nu$ is radiated in the direction described by the angle θ from the surface normal from infinitesimal surface area dA into infinitesimal solid angle $d\Omega$ in an infinitesimal frequency band of width $d\nu$ centered on frequency ν . The total power radiated into any solid angle is the integral of $B_\nu(\nu, T)$ over those three quantities, and is given by the Stefan–Boltzmann law. The spectral radiance of Planckian radiation from a black body has the same value for every direction and angle of polarization, and so the black body is said to be a Lambertian radiator.

Different forms

Planck's law can be encountered in several forms depending on the conventions and preferences of different scientific fields. The various forms of the law for spectral radiance are summarized in the table below. Forms on the left are most often encountered in experimental fields, while those on the right are most often encountered in theoretical fields.

Planck's law expressed in terms of different spectral variables^{[9][10][11]}

with h		with \hbar	
variable	distribution	variable	distribution
Frequency ν	$B_\nu(\nu, T) = \frac{2h\nu^3}{c^2} \frac{1}{e^{h\nu/(k_B T)} - 1}$	Angular frequency ω	$B_\omega(\omega, T) = \frac{\hbar\omega^3}{4\pi^3 c^2} \frac{1}{e^{\hbar\omega/(k_B T)} - 1}$
Wavelength λ	$B_\lambda(\lambda, T) = \frac{2hc^2}{\lambda^5} \frac{1}{e^{hc/(\lambda k_B T)} - 1}$	Angular wavelength y	$B_y(y, T) = \frac{\hbar c^2}{4\pi^3 y^5} \frac{1}{e^{\hbar c/(y k_B T)} - 1}$
Wavenumber $\tilde{\nu}$	$B_{\tilde{\nu}}(\tilde{\nu}, T) = 2hc^2 \tilde{\nu}^3 \frac{1}{e^{h c \tilde{\nu}/(k_B T)} - 1}$	Angular wavenumber k	$B_k(k, T) = \frac{\hbar c^2 k^3}{4\pi^3} \frac{1}{e^{\hbar ck/(k_B T)} - 1}$

These distributions represent the spectral radiance of blackbodies—the power emitted from the emitting surface, per unit projected area of emitting surface, per unit solid angle, per spectral unit (frequency, wavelength, wavenumber or their angular equivalents). Since the radiance is isotropic (i.e. independent of direction), the power emitted at an angle to the normal is proportional to the projected area, and therefore to the cosine of that angle as per Lambert's cosine law, and is unpolarized.

Correspondence between spectral variable forms

Different spectral variables require different corresponding forms of expression of the law. In general, one may not convert between the various forms of Planck's law simply by substituting one variable for another, because this would not take into account that the different forms have different units. Wavelength and frequency units are reciprocal.

Corresponding forms of expression are related because they express one and the same physical fact: for a particular physical spectral increment, a corresponding particular physical energy increment is radiated.

This is so whether it is expressed in terms of an increment of frequency, $d\nu$, or, correspondingly, of wavelength, $d\lambda$. Introduction of a minus sign can indicate that an increment of frequency corresponds with decrement of wavelength. For the above corresponding forms of expression of the spectral radiance, one may use an obvious expansion of notation, temporarily for the present calculation only. Then, for a particular spectral increment, the particular physical energy increment may be written

$$B_\lambda(\lambda, T) d\lambda = -B_\nu(\nu(\lambda), T) d\nu, \quad \text{which leads to} \quad B_\lambda(\lambda, T) = - \frac{d\nu}{d\lambda} B_\nu(\nu(\lambda), T).$$

Also, $\nu(\lambda) = \frac{c}{\lambda}$, so that $\frac{d\nu}{d\lambda} = -\frac{c}{\lambda^2}$. Substitution gives the correspondence between the frequency and wavelength forms, with their different dimensions and units.^{[11][12]} Consequently,

$$\frac{B_\lambda(T)}{B_\nu(T)} = \frac{c}{\lambda^2} = \frac{\nu^2}{c}.$$

Evidently, the location of the peak of the spectral distribution for Planck's law depends on the choice of spectral variable. Nevertheless, in a manner of speaking, this formula means that the shape of the spectral distribution is independent of temperature, according to Wien's displacement law, as detailed below in the sub-section **Percentiles** of the section Properties.

Spectral energy density form

Planck's law can also be written in terms of the spectral energy density (u) by multiplying B by $\frac{4\pi}{c}$:^[13]

$$u_i(T) = \frac{4\pi}{c} B_i(T).$$

These distributions have units of energy per volume per spectral unit.

First and second radiation constants

In the above variants of Planck's law, the *Wavelength* and *Wavenumber* variants use the terms $2hc^2$ and $\frac{hc}{k_B}$ which comprise physical constants only. Consequently, these terms can be considered as physical constants themselves,^[14] and are therefore referred to as the **first radiation constant** c_{1L} and the **second radiation constant** c_2 with

$$c_{1L} = 2hc^2$$

and

$$c_2 = \frac{hc}{k_B}.$$

Using the radiation constants, the *Wavelength* variant of Planck's law can be simplified to

$$L(\lambda, T) = \frac{c_{1L}}{\lambda^5} \frac{1}{\exp\left(\frac{c_2}{\lambda T}\right) - 1}$$

and the *wavenumber* variant can be simplified correspondingly.

L is used here instead of B because it is the SI symbol for *spectral radiance*. The L in c_{1L} refers to that. This reference is necessary because Planck's law can be reformulated to give spectral radiant exitance $M(\lambda, T)$ rather than *spectral radiance* $L(\lambda, T)$, in which case c_1 replaces c_{1L} , with

$$c_1 = 2\pi hc^2,$$

so that Planck's law for *spectral radiant exitance* can be written as

$$M(\lambda, T) = \frac{c_1}{\lambda^5} \frac{1}{\exp\left(\frac{c_2}{\lambda T}\right) - 1}$$

Derivation

Consider a cube of side L with conducting walls filled with electromagnetic radiation in thermal equilibrium at temperature T . If there is a small hole in one of the walls, the radiation emitted from the hole will be characteristic of a perfect black body. We will first calculate the spectral energy density within the cavity and then determine the spectral radiance of the emitted radiation.

At the walls of the cube, the parallel component of the electric field and the orthogonal component of the magnetic field must vanish. Analogous to the wave function of a particle in a box, one finds that the fields are superpositions of periodic functions. The three wavelengths λ_1 , λ_2 , and λ_3 , in the three directions orthogonal to the walls can be:

$$\lambda_i = \frac{2L}{n_i},$$

where the n_i are positive integers. For each set of integers n_i there are two linear independent solutions (modes). According to quantum theory, the energy levels of a mode are given by:

$$E_{n_1, n_2, n_3}(r) = \left(r + \frac{1}{2}\right) \frac{hc}{2L} \sqrt{n_1^2 + n_2^2 + n_3^2}. \quad (1)$$

The quantum number r can be interpreted as the number of photons in the mode. The two modes for each set of n_i correspond to the two polarization states of the photon which has a spin of 1. Note that for $r = 0$ the energy of the mode is not zero. This vacuum energy of the electromagnetic field is responsible for the Casimir effect. In the following we will calculate the internal energy of the box at absolute temperature T .

According to statistical mechanics, the probability distribution over the energy levels of a particular mode is given by:

$$P_r = \frac{\exp(-\beta E(r))}{Z(\beta)}.$$

Here

$$\beta \stackrel{\text{def}}{=} \frac{1}{k_B T}.$$

The denominator $Z(\beta)$, is the partition function of a single mode and makes P_r properly normalized:

$$Z(\beta) = \sum_{r=0}^{\infty} e^{-\beta E(r)} = \frac{e^{-\beta \epsilon/2}}{1 - e^{-\beta \epsilon}}.$$

Here we have implicitly defined

$$\epsilon \stackrel{\text{def}}{=} \frac{hc}{2L} \sqrt{n_1^2 + n_2^2 + n_3^2},$$

which is the energy of a single photon. As explained here, the average energy in a mode can be expressed in terms of the partition function:

$$\langle E \rangle = -\frac{d \log(Z)}{d\beta} = \frac{\epsilon}{2} + \frac{\epsilon}{e^{\beta \epsilon} - 1}.$$

This formula, apart from the first vacuum energy term, is a special case of the general formula for particles obeying Bose–Einstein statistics. Since there is no restriction on the total number of photons, the chemical potential is zero.

If we measure the energy relative to the ground state, the total energy in the box follows by summing $\langle E \rangle - \frac{\epsilon}{2}$ over all allowed single photon states. This can be done exactly in the thermodynamic limit as L approaches infinity. In this limit, ϵ becomes continuous and we can then integrate $\langle E \rangle - \frac{\epsilon}{2}$ over this parameter. To calculate the energy in the box in this way, we need to evaluate how many photon states there are in a given energy range. If we write the total number of single photon states with energies between ϵ and $\epsilon + d\epsilon$ as $g(\epsilon)d\epsilon$, where $g(\epsilon)$ is the density of states (which is evaluated below), then we can write:

$$U = \int_0^{\infty} \frac{\varepsilon}{e^{\beta\varepsilon} - 1} g(\varepsilon) d\varepsilon. \quad (2)$$

To calculate the density of states we rewrite equation (1) as follows:

$$\varepsilon \stackrel{\text{def}}{=} \frac{hc}{2L} n,$$

where n is the norm of the vector $\mathbf{n} = (n_1, n_2, n_3)$:

$$n = \sqrt{n_1^2 + n_2^2 + n_3^2}.$$

For every vector \mathbf{n} with integer components larger than or equal to zero, there are two photon states. This means that the number of photon states in a certain region of n -space is twice the volume of that region. An energy range of $d\varepsilon$ corresponds to shell of thickness $dn = \frac{2L}{hc} d\varepsilon$ in n -space. Because the components of \mathbf{n} have to be positive, this shell spans an octant of a sphere. The number of photon states $g(\varepsilon)d\varepsilon$, in an energy range $d\varepsilon$, is thus given by:

$$g(\varepsilon) d\varepsilon = 2 \frac{1}{8} 4\pi n^2 dn = \frac{8\pi L^3}{h^3 c^3} \varepsilon^2 d\varepsilon.$$

Inserting this in Eq. (2) gives:

$$U = L^3 \frac{8\pi}{h^3 c^3} \int_0^{\infty} \frac{\varepsilon^3}{e^{\beta\varepsilon} - 1} d\varepsilon. \quad (3)$$

From this equation one can derive the spectral energy density as a function of frequency $u_\nu(T)$ and as a function of wavelength $u_\lambda(T)$:

$$\frac{U}{L^3} = \int_0^{\infty} u_\nu(T) d\nu,$$

where

$$u_\nu(T) = \frac{8\pi h \nu^3}{c^3} \frac{1}{e^{h\nu/k_B T} - 1}.$$

And:

$$\frac{U}{L^3} = \int_0^{\infty} u_\lambda(T) d\lambda,$$

where

$$u_\lambda(T) = \frac{8\pi hc}{\lambda^5} \frac{1}{e^{hc/\lambda k_B T} - 1}.$$

This is also a spectral energy density function with units of energy per unit wavelength per unit volume. Integrals of this type for Bose and Fermi gases can be expressed in terms of polylogarithms. In this case, however, it is possible to calculate the integral in closed form using only elementary functions. Substituting

$$\varepsilon = k_B T x,$$

in Eq. (3), makes the integration variable dimensionless giving:

$$u(T) = \frac{8\pi (k_B T)^4}{(hc)^3} J,$$

where J is a Bose–Einstein integral given by:

$$J = \int_0^{\infty} \frac{x^3}{e^x - 1} dx = \frac{\pi^4}{15}.$$

The total electromagnetic energy inside the box is thus given by:

$$\frac{U}{V} = \frac{8\pi^5 (k_B T)^4}{15 (hc)^3},$$

where $V = L^3$ is the volume of the box.

The combination $\frac{hc}{k_B}$ has the value $14\,387.770\ \mu\text{m}\cdot\text{K}$.

This is **not** the Stefan–Boltzmann law (which provides the total energy *radiated* by a black body per unit surface area per unit time), but it can be written more compactly using the Stefan–Boltzmann constant σ , giving

$$\frac{U}{V} = \frac{4\sigma T^4}{c}.$$

The constant $\frac{4\sigma}{c}$ is sometimes called the radiation constant.

Since the radiation is the same in all directions, and propagates at the speed of light (c), the spectral radiance of radiation exiting the small hole is

$$B_\nu(T) = \frac{u_\nu(T) c}{4\pi},$$

which yields

$$B_\nu(T) = \frac{2h\nu^3}{c^2} \frac{1}{e^{h\nu/k_B T} - 1}.$$

It can be converted to an expression for $B_\lambda(T)$ in wavelength units by substituting ν by $\frac{c}{\lambda}$ and evaluating

$$B_\lambda(T) = B_\nu(T) \left| \frac{d\nu}{d\lambda} \right|.$$

Note that dimensional analysis shows that the unit of steradians, shown in the denominator of left hand side of the equation above, is generated in and carried through the derivation but does not appear in any of the dimensions for any element on the left-hand-side of the equation.

This derivation is based on Brehm & Mullin 1989.

Physics

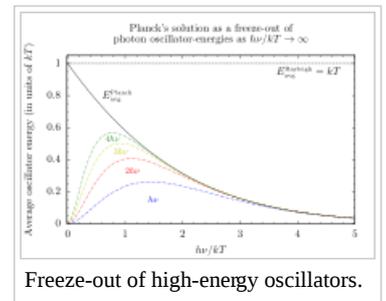
Outline

Planck's law describes the unique and characteristic spectral distribution for electromagnetic radiation in thermodynamic equilibrium, when there is no net flow of matter or energy.^[4] Its physics is most easily understood by considering the radiation in a cavity with rigid opaque walls. Motion of the walls can affect the radiation. If the walls are not opaque, then the thermodynamic equilibrium is not isolated. It is of interest to explain how the thermodynamic equilibrium is attained. There are two main cases: (a) when the approach to thermodynamic equilibrium is in the presence of matter, when the walls of the cavity are imperfectly reflective for every wavelength or when the walls are perfectly reflective while the cavity contains a small black body (this was the main case considered by Planck); or (b) when the approach to equilibrium is in the absence of matter, when the walls are perfectly reflective for all wavelengths and the cavity contains no matter. For matter not enclosed in such a cavity, thermal radiation can be approximately explained by appropriate use of Planck's law.

Classical physics led, via the Equipartition theorem, to the Ultraviolet catastrophe, a prediction that the total blackbody radiation intensity was infinite. If supplemented by the classically unjustifiable assumption that for some reason the radiation is finite, classical thermodynamics provides an account of some aspects of the Planck distribution, such as the Stefan–Boltzmann law, and the Wien displacement law. For the case of the presence of matter, quantum mechanics provides a good account, as found below in the section headed Einstein coefficients. This was the case considered by Einstein, and is nowadays used for quantum optics.^{[15][16]} For the case of the absence of matter, quantum field theory is necessary, because non-relativistic quantum mechanics with fixed particle numbers does not provide a sufficient account.

Photons

Quantum theoretical explanation of Planck's law views the radiation as a gas of massless, uncharged, bosonic particles, namely photons, in thermodynamic equilibrium. Photons are viewed as the carriers of the electromagnetic interaction between electrically charged elementary particles. Photon numbers are not conserved. Photons are created or annihilated in the right numbers and with the right energies to fill the cavity with the Planck distribution. For a photon gas in thermodynamic equilibrium, the internal energy density is entirely determined by the temperature; moreover, the pressure is entirely determined by the internal energy density. This is unlike the case of thermodynamic equilibrium for material gases, for which the internal energy is determined not only by the temperature, but also, independently, by the



respective numbers of the different molecules, and independently again, by the specific characteristics of the different molecules. For different material gases at given temperature, the pressure and internal energy density can vary independently, because different molecules can carry independently different excitation energies.

Planck's law arises as a limit of the Bose–Einstein distribution, the energy distribution describing non-interactive bosons in thermodynamic equilibrium. In the case of massless bosons such as photons and gluons, the chemical potential is zero and the Bose–Einstein distribution reduces to the Planck distribution. There is another fundamental equilibrium energy distribution: the Fermi–Dirac distribution, which describes fermions, such as electrons, in thermal equilibrium. The two distributions differ because multiple bosons can occupy the same quantum state, while multiple fermions cannot. At low densities, the number of available quantum states per particle is large, and this difference becomes irrelevant. In the low density limit, the Bose-Einstein and the Fermi-Dirac distribution each reduce to the Maxwell–Boltzmann distribution.

Kirchhoff's law of thermal radiation

Kirchhoff's law of thermal radiation is a succinct and brief account of a complicated physical situation. The following is an introductory sketch of that situation, and is very far from being a rigorous physical argument. The purpose here is only to summarize the main physical factors in the situation, and the main conclusions.

Spectral dependence of thermal radiation

There is a difference between conductive heat transfer and radiative heat transfer. Radiative heat transfer can be filtered to pass only a definite band of radiative frequencies.

It is generally known that the hotter a body becomes the more heat it radiates at every frequency.

In a cavity in an opaque body with rigid walls that are not perfectly reflective at any frequency, in thermodynamic equilibrium, there is only one temperature, and it must be shared in common by the radiation of every frequency.

One may imagine two such cavities, each in its own isolated radiative and thermodynamic equilibrium. One may imagine an optical device that allows radiative heat transfer between the two cavities, filtered to pass only a definite band of radiative frequencies. If the values of the spectral radiances of the radiations in the cavities differ in that frequency band, heat may be expected to pass from the hotter to the colder. One might propose to use such a filtered transfer of heat in such a band to drive a heat engine. If the two bodies are at the same temperature, the second law of thermodynamics does not allow the heat engine to work. It may be inferred that for a temperature common to the two bodies, the values of the spectral radiances in the pass-band must also be common. This must hold for every frequency band.^{[17][18][19]} This became clear to Balfour Stewart and later to Kirchhoff. Balfour Stewart found experimentally that of all surfaces, one of lamp-black emitted the greatest amount of thermal radiation for every quality of radiation, judged by various filters.

Thinking theoretically, Kirchhoff went a little further, and pointed out that this implied that the spectral radiance, as a function of radiative frequency, of any such cavity in thermodynamic equilibrium must be a unique universal function of temperature. He postulated an ideal black body that interfaced with its surrounds in just such a way as to absorb all the radiation that falls on it. By the Helmholtz reciprocity principle, radiation from the interior of such a body would pass unimpeded, directly to its surrounds without reflection at the interface. In thermodynamic equilibrium, the thermal radiation emitted from such a body would have that unique universal spectral radiance as a function of temperature. This insight is the root of Kirchhoff's law of thermal radiation.

Relation between absorptivity and emissivity

One may imagine a small homogeneous spherical material body labeled X at a temperature T_X , lying in a radiation field within a large cavity with walls of material labeled Y at a temperature T_Y . The body X emits its own thermal radiation. At a particular frequency ν , the radiation emitted from a particular cross-section through the centre of X in one sense in a direction normal to that cross-section may be denoted $I_{\nu,X}(T_X)$, characteristically for the material of X . At that frequency ν , the radiative power from the walls into that cross-section in the opposite sense in that direction may be denoted $I_{\nu,Y}(T_Y)$, for the wall temperature T_Y . For the material of X , defining the absorptivity $\alpha_{\nu,X,Y}(T_X, T_Y)$ as the fraction of that incident radiation absorbed by X , that incident energy is absorbed at a rate $\alpha_{\nu,X,Y}(T_X, T_Y) I_{\nu,Y}(T_Y)$.

The rate $q(\nu, T_X, T_Y)$ of accumulation of energy in one sense into the cross-section of the body can then be expressed

$$q(\nu, T_X, T_Y) = \alpha_{\nu,X,Y}(T_X, T_Y) I_{\nu,Y}(T_Y) - I_{\nu,X}(T_X).$$

Kirchhoff's seminal insight, mentioned just above, was that, at thermodynamic equilibrium at temperature T , there exists a unique universal radiative distribution, nowadays denoted $B_\nu(T)$, that is independent of the chemical characteristics of the materials X and Y , that leads to a very valuable understanding of the radiative exchange equilibrium of any body at all, as follows.

When there is thermodynamic equilibrium at temperature T , the cavity radiation from the walls has that unique universal value, so that $I_{\nu,Y}(T_Y) = B_\nu(T)$. Further, one may define the emissivity $\varepsilon_{\nu,X}(T_X)$ of the material of the body X just so that at thermodynamic equilibrium at temperature $T_X = T$, one has $I_{\nu,X}(T_X) = I_{\nu,X}(T) = \varepsilon_{\nu,X}(T) B_\nu(T)$.

When thermal equilibrium prevails at temperature $T = T_X = T_Y$, the rate of accumulation of energy vanishes so that $q(\nu, T_X, T_Y) = 0$. It follows that in thermodynamic equilibrium, when $T = T_X = T_Y$,

$$0 = \alpha_{\nu,X,Y}(T,T)B_{\nu}(T) - \epsilon_{\nu,X}(T)B_{\nu}(T).$$

Kirchhoff pointed out that it follows that in thermodynamic equilibrium, when $T = T_X = T_Y$,

$$\alpha_{\nu,X,Y}(T,T) = \epsilon_{\nu,X}(T).$$

Introducing the special notation $\alpha_{\nu,X}(T)$ for the absorptivity of material X at thermodynamic equilibrium at temperature T (justified by a discovery of Einstein, as indicated below), one further has the equality

$$\alpha_{\nu,X}(T) = \epsilon_{\nu,X}(T)$$

at thermodynamic equilibrium.

The equality of absorptivity and emissivity here demonstrated is specific for thermodynamic equilibrium at temperature T and is in general not to be expected to hold when conditions of thermodynamic equilibrium do not hold. The emissivity and absorptivity are each separately properties of the molecules of the material but they depend differently upon the distributions of states of molecular excitation on the occasion, because of a phenomenon known as "stimulated emission", that was discovered by Einstein. On occasions when the material is in thermodynamic equilibrium or in a state known as local thermodynamic equilibrium, the emissivity and absorptivity become equal. Very strong incident radiation or other factors can disrupt thermodynamic equilibrium or local thermodynamic equilibrium. Local thermodynamic equilibrium in a gas means that molecular collisions far outweigh light emission and absorption in determining the distributions of states of molecular excitation.

Kirchhoff pointed out that he did not know the precise character of $B_{\nu}(T)$, but he thought it important that it should be found out. Four decades after Kirchhoff's insight of the general principles of its existence and character, Planck's contribution was to determine the precise mathematical expression of that equilibrium distribution $B_{\nu}(T)$.

Black body

In physics, one considers an ideal black body, here labeled B , defined as one that completely absorbs all of the electromagnetic radiation falling upon it at every frequency ν (hence the term "black"). According to Kirchhoff's law of thermal radiation, this entails that, for every frequency ν , at thermodynamic equilibrium at temperature T , one has $\alpha_{\nu,B}(T) = \epsilon_{\nu,B}(T) = 1$, so that the thermal radiation from a black body is always equal to the full amount specified by Planck's law. No physical body can emit thermal radiation that exceeds that of a black body, since if it were in equilibrium with a radiation field, it would be emitting more energy than was incident upon it.

Though perfectly black materials do not exist, in practice a black surface can be accurately approximated.^[4] As to its material interior, a body of condensed matter, liquid, solid, or plasma, with a definite interface with its surroundings, is completely black to radiation if it is completely opaque. That means that it absorbs all of the radiation that penetrates the interface of the body with its surroundings, and enters the body. This is not too difficult to achieve in practice. On the other hand, a perfectly black interface is not found in nature. A perfectly black interface reflects no radiation, but transmits all that falls on it, from either side. The best practical way to make an effectively black interface is to simulate an 'interface' by a small hole in the wall of a large cavity in a completely opaque rigid body of material that does not reflect perfectly at any frequency, with its walls at a controlled temperature. Beyond these requirements, the component material of the walls is unrestricted. Radiation entering the hole has almost no possibility of escaping the cavity without being absorbed by multiple impacts with its walls.^[20]

Lambert's cosine law

As explained by Planck,^[21] a radiating body has an interior consisting of matter, and an interface with its contiguous neighbouring material medium, which is usually the medium from within which the radiation from the surface of the body is observed. The interface is not composed of physical matter but is a theoretical conception, a mathematical two-dimensional surface, a joint property of the two contiguous media, strictly speaking belonging to neither separately. Such an interface can neither absorb nor emit, because it is not composed of physical matter; but it is the site of reflection and transmission of radiation, because it is a surface of discontinuity of optical properties. The reflection and transmission of radiation at the interface obey the Stokes–Helmholtz reciprocity principle.

At any point in the interior of a black body located inside a cavity in thermodynamic equilibrium at temperature T the radiation is homogeneous, isotropic and unpolarized. A black body absorbs all and reflects none of the electromagnetic radiation incident upon it. According to the Helmholtz reciprocity principle, radiation from the interior of a black body is not reflected at its surface, but is fully transmitted to its exterior. Because of the isotropy of the radiation in the body's interior, the spectral radiance of radiation transmitted from its interior to its exterior through its surface is independent of direction.^[22]

This is expressed by saying that radiation from the surface of a black body in thermodynamic equilibrium obeys Lambert's cosine law.^{[23][24]} This means that the spectral flux $d\Phi(dA, \theta, d\Omega, d\nu)$ from a given infinitesimal element of area dA of the actual emitting surface of the black body, detected from a given direction that makes an angle θ with the normal to the actual emitting surface at dA , into an element of solid angle of detection $d\Omega$ centred on the direction indicated by θ , in an element of frequency bandwidth $d\nu$, can be represented as^[25]

$$\frac{d\Phi(dA, \theta, d\Omega, d\nu)}{d\Omega} = L^0(dA, d\nu) dA d\nu \cos \theta$$

where $L^0(dA, d\nu)$ denotes the flux, per unit area per unit frequency per unit solid angle, that area dA would show if it were measured in its normal direction $\theta = 0$.

The factor $\cos \theta$ is present because the area to which the spectral radiance refers directly is the projection, of the actual emitting surface area, onto a plane perpendicular to the direction indicated by θ . This is the reason for the name *cosine law*.

Taking into account the independence of direction of the spectral radiance of radiation from the surface of a black body in thermodynamic equilibrium, one has $L^0(dA, d\nu) = B_\nu(T)$ and so

$$\frac{d\Phi(dA, \theta, d\Omega, d\nu)}{d\Omega} = B_\nu(T) dA d\nu \cos \theta.$$

Thus Lambert's cosine law expresses the independence of direction of the spectral radiance $B_\nu(T)$ of the surface of a black body in thermodynamic equilibrium.

Stefan–Boltzmann law

The total power emitted per unit area at the surface of a black body (P) may be found by integrating the black body spectral flux found from Lambert's law over all frequencies, and over the solid angles corresponding to a hemisphere (h) above the surface.

$$P = \int_0^\infty d\nu \int_h d\Omega B_\nu \cos(\theta)$$

The infinitesimal solid angle can be expressed in spherical polar coordinates:

$$d\Omega = \sin(\theta) d\theta d\phi.$$

So that:

$$P = \int_0^\infty d\nu \int_0^{\frac{\pi}{2}} d\theta \int_0^{2\pi} d\phi B_\nu(T) \cos(\theta) \sin(\theta) = \sigma T^4$$

where

$$\sigma = \frac{2k_B^4 \pi^5}{15c^2 h^3} \approx 5.670400 \times 10^{-8} \text{ J s}^{-1} \text{ m}^{-2} \text{ K}^{-4}$$

is known as the Stefan–Boltzmann constant.^[26]

Radiative transfer

The equation of radiative transfer describes the way in which radiation is affected as it travels through a material medium. For the special case in which the material medium is in thermodynamic equilibrium in the neighborhood of a point in the medium, Planck's law is of special importance.

For simplicity, we can consider the linear steady state, without scattering. The equation of radiative transfer states that for a beam of light going through a small distance ds , energy is conserved: The change in the (spectral) radiance of that beam (I_ν) is equal to the amount removed by the material medium plus the amount gained from the material medium. If the radiation field is in equilibrium with the material medium, these two contributions will be equal. The material medium will have a certain emission coefficient and absorption coefficient.

The absorption coefficient α is the fractional change in the intensity of the light beam as it travels the distance ds , and has units of length^{-1} . It is composed of two parts, the decrease due to absorption and the increase due to stimulated emission. Stimulated emission is emission by the material body which is caused by and is proportional to the incoming radiation. It is included in the absorption term because, like absorption, it is proportional to the intensity of the incoming radiation. Since the amount of absorption will generally vary linearly as the density ρ of the material, we may define a "mass absorption coefficient" $\kappa_\nu = \frac{\alpha}{\rho}$ which is a property of the material itself. The change in intensity of a light beam due to absorption as it traverses a small distance ds will then be^[2]

$$dI_\nu = -\kappa_\nu \rho I_\nu ds$$

The "mass emission coefficient" j_ν is equal to the radiance per unit volume of a small volume element divided by its mass (since, as for the mass absorption coefficient, the emission is proportional to the emitting mass) and has units of $\text{power} \cdot \text{solid angle}^{-1} \cdot \text{frequency}^{-1} \cdot \text{density}^{-1}$. Like the mass absorption coefficient, it too is a property of the material itself. The change in a light beam as it traverses a small distance ds will then be^[27]

$$dI_\nu = j_\nu \rho ds$$

The equation of radiative transfer will then be the sum of these two contributions:^[28]

$$\frac{dI_\nu}{ds} = j_\nu \rho - \kappa_\nu \rho I_\nu.$$

If the radiation field is in equilibrium with the material medium, then the radiation will be homogeneous (independent of position) so that $dI_\nu = 0$ and:

$$\kappa_\nu B_\nu = j_\nu$$

which is another statement of Kirchhoff's law, relating two material properties of the medium, and which yields the radiative transfer equation at a point around which the medium is in thermodynamic equilibrium:

$$\frac{dI_\nu}{ds} = \kappa_\nu \rho (B_\nu - I_\nu).$$

Einstein coefficients

The principle of detailed balance states that, at thermodynamic equilibrium, each elementary process is equilibrated by its reverse process.

In 1916, Albert Einstein applied this principle on an atomic level to the case of an atom radiating and absorbing radiation due to transitions between two particular energy levels,^[29] giving a deeper insight into the equation of radiative transfer and Kirchhoff's law for this type of radiation. If level 1 is the lower energy level with energy E_1 , and level 2 is the upper energy level with energy E_2 , then the frequency ν of the radiation radiated or absorbed will be determined by Bohr's frequency condition:^{[30][31]}

$$E_2 - E_1 = h\nu.$$

If n_1 and n_2 are the number densities of the atom in states 1 and 2 respectively, then the rate of change of these densities in time will be due to three processes:

$$\left(\frac{dn_1}{dt} \right)_{\text{spon}} = A_{21} n_2 \quad \text{Spontaneous emission}$$

$$\left(\frac{dn_1}{dt} \right)_{\text{stim}} = B_{21} n_2 I_\nu(T) \quad \text{Stimulated emission}$$

$$\left(\frac{dn_2}{dt} \right)_{\text{abs}} = B_{12} n_1 I_\nu(T) \quad \text{Photo-absorption}$$

where $I_\nu(T)$ is the spectral radiance of the radiation field. The three parameters A_{21} , B_{21} and B_{12} , known as the Einstein coefficients, are associated with the photon frequency ν produced by the transition between two energy levels (states). As a result, each line in a spectra has its own set of associated coefficients. When the atoms and the radiation field are in equilibrium, the radiance will be given by Planck's law and, by the principle of detailed balance, the sum of these rates must be zero:

$$0 = A_{21} n_2 + B_{21} n_2 B_\nu(T) - B_{12} n_1 B_\nu(T)$$

Since the atoms are also in equilibrium, the populations of the two levels are related by the Boltzmann factor:

$$\frac{n_2}{n_1} = \frac{g_2}{g_1} e^{-h\nu/k_B T}$$

where g_1 and g_2 are the multiplicities of the respective energy levels. Combining the above two equations with the requirement that they be valid at any temperature yields two relationships between the Einstein coefficients:

$$\frac{A_{21}}{B_{21}} = \frac{2h\nu^3}{c^2}$$

$$\frac{B_{21}}{B_{12}} = \frac{g_1}{g_2}$$

so that knowledge of one coefficient will yield the other two. For the case of isotropic absorption and emission, the emission coefficient (j_ν) and absorption coefficient (κ_ν) defined in the radiative transfer section above, can be expressed in terms of the Einstein coefficients. The relationships between the Einstein coefficients will yield the expression of Kirchhoff's law expressed in the *Radiative transfer* section above, namely that

$$j_\nu = \kappa_\nu B_\nu.$$

These coefficients apply to both atoms and molecules.

Properties

Peaks

The distributions B_ν , B_ω , $B_{\tilde{\nu}}$ and B_k peak at a photon energy of^[32]

$$E = \left[3 + W\left(\frac{-3}{e^3}\right) \right] k_B T \approx 2.821 k_B T,$$

where W is the Lambert W function and e is Euler's number.

The distributions B_λ and B_γ however, peak at a different energy^[32]

$$E = \left[5 + W\left(\frac{-5}{e^5}\right) \right] k_B T \approx 4.965 k_B T,$$

The reason for this is that, as mentioned above, one cannot go from (for example) B_ν to B_λ simply by substituting ν by λ . In addition, one must also multiply the result of the substitution by

$$\left| \frac{d\nu}{d\lambda} \right| = c/\lambda^2.$$

This $\frac{1}{\lambda^2}$ factor shifts the peak of the distribution to higher energies.

Dividing hc by this energy expression gives the wavelength of the peak. For this one can use $\frac{hc}{k_B} = 14\,387.770 \mu\text{m}\cdot\text{K}$.

The spectral radiance at these peaks is given by:

$$B_{\nu,\max}(T) = \frac{2k_B^3 T^3 (3 + W(-3 \exp(-3)))^3}{h^2 c^2} \frac{1}{e^{3+W(-3 \exp(-3))} - 1} \approx \left(1.896 \times 10^{-19} \frac{\text{W}}{\text{m}^2 \cdot \text{Hz} \cdot \text{K}^3} \right) \times T^3$$

$$B_{\lambda,\max}(T) = \frac{2k_B^5 T^5 (5 + W(-5 \exp(-5)))^5}{h^4 c^3} \frac{1}{e^{5+W(-5 \exp(-5))} - 1} \approx \left(4.096 \times 10^{-6} \frac{\text{W}}{\text{m}^3 \cdot \text{K}^5} \right) \times T^5$$

Approximations

In the limit of low frequencies (i.e. long wavelengths), Planck's law becomes the Rayleigh–Jeans law^{[33][34][35]}

$$B_\nu(T) \approx \frac{2\nu^2}{c^2} k_B T \quad \text{or} \quad B_\lambda(T) \approx \frac{2c}{\lambda^4} k_B T.$$

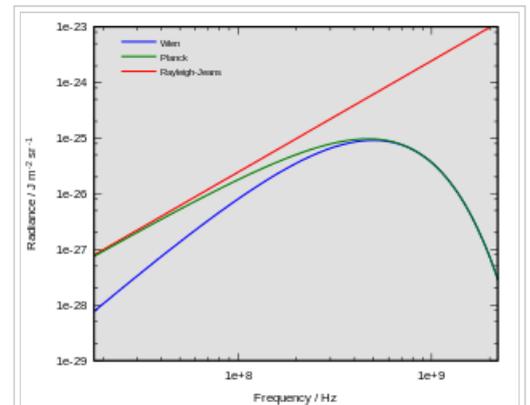
The radiance increases as the square of the frequency, illustrating the ultraviolet catastrophe. In the limit of high frequencies (i.e. small wavelengths) Planck's law tends to the Wien approximation:^{[35][36][37]}

$$B_\nu(T) \approx \frac{2h\nu^3}{c^2} e^{-\frac{h\nu}{k_B T}} \quad \text{or} \quad B_\lambda(T) \approx \frac{2hc^2}{\lambda^5} e^{-\frac{hc}{\lambda k_B T}}.$$

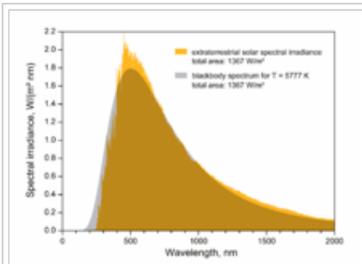
Both approximations were known to Planck before he developed his law. He was led by these two approximations to develop a law which incorporated both limits, which ultimately became Planck's law.

Percentiles

Wien's displacement law in its stronger form states that the shape of Planck's law is independent of temperature. It is therefore possible to list the percentile points of the total radiation as well as the peaks for wavelength and frequency, in a form which gives the wavelength λ when divided by temperature T .^[38] The second row of the following table lists the corresponding values of λT , that is, those values of x for which the wavelength λ is $\frac{x}{T}$ micrometers at the radiance percentile point given by the corresponding entry in the first row.



Log-log plots of radiance vs. frequency for Planck's law (green), compared with the Rayleigh–Jeans law (red) and the Wien approximation (blue) for a black body at 8 mK temperature.



The Sun is an excellent approximation of a black body. Its effective temperature is ~5777 K.

Percentile	0.01%	0.1%	1%	10%	20%	25.0%	30%	40%	41.8%	50%	60%	64.6%	70%	80%	90%	99%	99.9%	99.99%
λT ($\mu\text{m}\cdot\text{K}$)	910	1110	1448	2195	2676	2898	3119	3582	3670	4107	4745	5099	5590	6864	9376	22884	51613	113374

That is, 0.01% of the radiation is at a wavelength below $\frac{910}{T}$ μm , 20% below $\frac{2676}{T}$ μm , etc. The wavelength and frequency peaks are in bold and occur at 25.0% and 64.6% respectively. The 41.8% point is the wavelength-frequency-neutral peak. These are the points at which the respective Planck-law functions $\frac{1}{\lambda^5}$, ν^3 and $\frac{\nu^2}{\lambda^2}$ divided by $\exp\left(\frac{h\nu}{k_B T}\right) - 1$ attain their maxima. Also note the much smaller gap in ratio of wavelengths between 0.1% and 0.01% (1110 is 22% more than 910) than between 99.9% and 99.99% (113374 is 120% more than 51613), reflecting the exponential decay of energy at short wavelengths (left end) and polynomial decay at long.

Which peak to use depends on the application. The conventional choice is the wavelength peak at 25.0% given by Wien's displacement law in its weak form. For some purposes the median or 50% point dividing the total radiation into two halves may be more suitable. The latter is closer to the frequency peak than to the wavelength peak because the radiance drops exponentially at short wavelengths and only polynomially at long. The neutral peak occurs at a shorter wavelength than the median for the same reason.

For the Sun, T is 5778 K, allowing the percentile points of the Sun's radiation, in nanometers, to be tabulated as follows when modeled as a black body radiator, to which the Sun is a fair approximation. For comparison a planet modeled as a black body radiating at a nominal 288 K (15 °C) as a representative value of the Earth's highly variable temperature has wavelengths more than twenty times that of the Sun, tabulated in the third row in micrometers (thousands of nanometers).

Percentile	0.01%	0.1%	1%	10%	20%	25.0%	30%	40%	41.8%	50%	60%	64.6%	70%	80%	90%	99%	99.9%	99.99%
Sun λ (μm)	0.157	0.192	0.251	0.380	0.463	0.502	0.540	0.620	0.635	0.711	0.821	0.882	0.967	1.188	1.623	3.961	8.933	19.620
288 K planet λ (μm)	3.16	3.85	5.03	7.62	9.29	10.1	10.8	12.4	12.7	14.3	16.5	17.7	19.4	23.8	32.6	79.5	179	394

That is, only 1% of the Sun's radiation is at wavelengths shorter than 251 nm, and only 1% at longer than 3961 nm. Expressed in micrometers this puts 98% of the Sun's radiation in the range from 0.251 to 3.961 μm . The corresponding 98% of energy radiated from a 288 K planet is from 5.03 to 79.5 μm , well above the range of solar radiation (or below if expressed in terms of frequencies $\nu = \frac{c}{\lambda}$ instead of wavelengths λ).

A consequence of this more-than-order-of-magnitude difference in wavelength between solar and planetary radiation is that filters designed to pass one and block the other are easy to construct. For example, windows fabricated of ordinary glass or transparent plastic pass at least 80% of the incoming 5778 K solar radiation, which is below 1.2 μm in wavelength, while blocking over 99% of the outgoing 288 K thermal radiation from 5 μm upwards, wavelengths at which most kinds of glass and plastic of construction-grade thickness are effectively opaque.

The Sun's radiation is that arriving at the top of the atmosphere (TOA). As can be read from the table, radiation below 400 nm, or ultraviolet, is about 12%, while that above 700 nm, or infrared, starts at about the 49% point and so accounts for 51% of the total. Hence only 37% of the TOA insolation is visible to the human eye. The atmosphere shifts these percentages substantially in favor of visible light as it absorbs most of the ultraviolet and significant amounts of infrared.

History

Forerunners

Balfour Stewart

In 1858, Balfour Stewart described his experiments on the thermal radiative emissive and absorptive powers of polished plates of various substances, compared with the powers of lamp-black surfaces, at the same temperature.^[5] Stewart chose lamp-black surfaces as his reference because of various previous experimental findings, especially those of Pierre Prevost and of John Leslie. He wrote "Lamp-black,

which absorbs all the rays that fall upon it, and therefore possesses the greatest possible absorbing power, will possess also the greatest possible radiating power."

Stewart measured radiated power with a thermo-pile and sensitive galvanometer read with a microscope. He was concerned with selective thermal radiation, which he investigated with plates of substances that radiated and absorbed selectively for different qualities of radiation rather than maximally for all qualities of radiation. He discussed the experiments in terms of rays which could be reflected and refracted, and which obeyed the Helmholtz reciprocity principle (though he did not use an eponym for it). He did not in this paper mention that the qualities of the rays might be described by their wavelengths, nor did he use spectrally resolving apparatus such as prisms or diffraction gratings. His work was quantitative within these constraints. He made his measurements in a room temperature environment, and quickly so as to catch his bodies in a condition near the thermal equilibrium in which they had been prepared by heating to equilibrium with boiling water. His measurements confirmed that substances that emit and absorb selectively respect the principle of selective equality of emission and absorption at thermal equilibrium.

Stewart offered a theoretical proof that this should be the case separately for every selected quality of thermal radiation, but his mathematics was not rigorously valid. According to historian D. M. Siegel: "He was not a practitioner of the more sophisticated techniques of nineteenth-century mathematical physics; he did not even make use of the functional notation in dealing with spectral distributions."^[39] He made no mention of thermodynamics in this paper, though he did refer to conservation of *vis viva*. He proposed that his measurements implied that radiation was both absorbed and emitted by particles of matter throughout depths of the media in which it propagated. He applied the Helmholtz reciprocity principle to account for the material interface processes as distinct from the processes in the interior material. He concluded that his experiments showed that, in the interior of an enclosure in thermal equilibrium, the radiant heat, reflected and emitted combined, leaving any part of the surface, regardless of its substance, was the same as would have left that same portion of the surface if it had been composed of lamp-black. He did not mention the possibility of ideally perfectly reflective walls; in particular he noted that highly polished real physical metals absorb very slightly.

Gustav Kirchhoff

In 1859, not knowing of Stewart's work, Gustav Robert Kirchhoff reported the coincidence of the wavelengths of spectrally resolved lines of absorption and of emission of visible light. Importantly for thermal physics, he also observed that bright lines or dark lines were apparent depending on the temperature difference between emitter and absorber.^[40]

Kirchhoff then went on to consider bodies that emit and absorb heat radiation, in an opaque enclosure or cavity, in equilibrium at temperature T .

Here is used a notation different from Kirchhoff's. Here, the emitting power $E(T, i)$ denotes a dimensioned quantity, the total radiation emitted by a body labeled by index i at temperature T . The total absorption ratio $a(T, i)$ of that body is dimensionless, the ratio of absorbed to incident radiation in the cavity at temperature T . (In contrast with Balfour Stewart's, Kirchhoff's definition of his absorption ratio did not refer in particular to a lamp-black surface as the source of the incident radiation.) Thus the ratio $\frac{E(T, i)}{a(T, i)}$ of emitting power to absorption ratio is a dimensioned quantity, with the dimensions of emitting power, because $a(T, i)$ is dimensionless. Also here the wavelength-specific emitting power of the body at temperature T is denoted by $E(\lambda, T, i)$ and the wavelength-specific absorption ratio by $a(\lambda, T, i)$. Again, the ratio $\frac{E(\lambda, T, i)}{a(\lambda, T, i)}$ of emitting power to absorption ratio is a dimensioned quantity, with the dimensions of emitting power.

In a second report made in 1859, Kirchhoff announced a new general principle or law for which he offered a theoretical and mathematical proof, though he did not offer quantitative measurements of radiation powers.^[41] His theoretical proof was and still is considered by some writers to be invalid.^{[39][42]} His principle, however, has endured: it was that for heat rays of the same wavelength, in equilibrium at a given temperature, the wavelength-specific ratio of emitting power to absorption ratio has one and the same common value for all bodies that emit and absorb at that wavelength. In symbols, the law stated that the wavelength-specific ratio $\frac{E(\lambda, T, i)}{a(\lambda, T, i)}$ has one and the same value for all bodies, that is for all values of index i . In this report there was no mention of black bodies.

In 1860, still not knowing of Stewart's measurements for selected qualities of radiation, Kirchhoff pointed out that it was long established experimentally that for total heat radiation, of unselected quality, emitted and absorbed by a body in equilibrium, the dimensioned total radiation ratio $\frac{E(T, i)}{a(T, i)}$, has one and the same value common to all bodies, that is, for every value of the material index i .^[43] Again without measurements of radiative powers or other new experimental data, Kirchhoff then offered a fresh theoretical proof of his new principle of the universality of the value of the wavelength-specific ratio $\frac{E(\lambda, T, i)}{a(\lambda, T, i)}$ at thermal equilibrium. His fresh theoretical proof was and still is considered by some writers to be invalid.^{[39][42]}

But more importantly, it relied on a new theoretical postulate of "**perfectly black bodies**", which is the reason why one speaks of Kirchhoff's law. Such black bodies showed complete absorption in their infinitely thin most superficial surface. They correspond to Balfour Stewart's reference bodies, with internal radiation, coated with lamp-black. They were not the more realistic perfectly black bodies later considered by Planck. Planck's black bodies radiated and absorbed only by the material in their interiors; their interfaces with contiguous media were only mathematical surfaces, capable neither of absorption nor emission, but only of reflecting and transmitting with refraction.^[44]

Kirchhoff's proof considered an arbitrary non-ideal body labeled i as well as various perfect black bodies labeled BB . It required that the bodies be kept in a cavity in thermal equilibrium at temperature T . His proof intended to show that the ratio $\frac{E(\lambda, T, i)}{a(\lambda, T, i)}$ was independent of the nature i of the non-ideal body, however partly transparent or partly reflective it was.

His proof first argued that for wavelength λ and at temperature T , at thermal equilibrium, all perfectly black bodies of the same size and shape have the one and the same common value of emissive power $E(\lambda, T, BB)$, with the dimensions of power. His proof noted that the dimensionless wavelength-specific absorption ratio $a(\lambda, T, BB)$ of a perfectly black body is by definition exactly 1. Then for a perfectly black body, the wavelength-specific ratio of emissive power to absorption ratio $\frac{E(\lambda, T, BB)}{a(\lambda, T, BB)}$ is again just $E(\lambda, T, BB)$, with the dimensions of power. Kirchhoff considered, successively, thermal equilibrium with the arbitrary non-ideal body, and with a perfectly black body of the same size and shape, in place in his cavity in equilibrium at temperature T . He argued that the flows of heat radiation must be the same in each case. Thus he argued that at thermal equilibrium the ratio $\frac{E(\lambda, T, i)}{a(\lambda, T, i)}$ was equal to $E(\lambda, T, BB)$, which may now be denoted $B_\lambda(\lambda, T)$, a continuous function, dependent only on λ at fixed temperature T , and an increasing function of T at fixed wavelength λ , at low temperatures vanishing for visible but not for longer wavelengths, with positive values for visible wavelengths at higher temperatures, which does not depend on the nature i of the arbitrary non-ideal body. (Geometrical factors, taken into detailed account by Kirchhoff, have been ignored in the foregoing.)

Thus **Kirchhoff's law of thermal radiation** can be stated: *For any material at all, radiating and absorbing in thermodynamic equilibrium at any given temperature T , for every wavelength λ , the ratio of emissive power to absorptive ratio has one universal value, which is characteristic of a perfect black body, and is an emissive power which we here represent by $B_\lambda(\lambda, T)$.* (For our notation $B_\lambda(\lambda, T)$, Kirchhoff's original notation was simply e .)^{[2][43][45][46][47][48]}

Kirchhoff announced that the determination of the function $B_\lambda(\lambda, T)$ was a problem of the highest importance, though he recognized that there would be experimental difficulties to be overcome. He supposed that like other functions that do not depend on the properties of individual bodies, it would be a simple function. That function $B_\lambda(\lambda, T)$ has occasionally been called 'Kirchhoff's (emission, universal) function',^{[49][50][51][52]} though its precise mathematical form would not be known for another forty years, till it was discovered by Planck in 1900. The theoretical proof for Kirchhoff's universality principle was worked on and debated by various physicists over the same time, and later.^[42] Kirchhoff stated later in 1860 that his theoretical proof was better than Balfour Stewart's, and in some respects it was so.^[39] Kirchhoff's 1860 paper did not mention the second law of thermodynamics, and of course did not mention the concept of entropy which had not at that time been established. In a more considered account in a book in 1862, Kirchhoff mentioned the connection of his law with "Carnot's principle", which is a form of the second law.^[53]

According to Helge Kragh, "Quantum theory owes its origin to the study of thermal radiation, in particular to the "blackbody" radiation that Robert Kirchhoff had first defined in 1859–1860."^[54]

Empirical and theoretical ingredients for the scientific induction of Planck's law

In 1860, Kirchhoff predicted experimental difficulties for the empirical determination of the function that described the dependence of the black-body spectrum as a function only of temperature and wavelength. And so it turned out. It took some forty years of development of improved methods of measurement of electromagnetic radiation to get a reliable result.^[55]

In 1865, John Tyndall described radiation from electrically heated filaments and from carbon arcs as visible and invisible.^[56] Tyndall spectrally decomposed the radiation by use of a rock salt prism, which passed heat as well as visible rays, and measured the radiation intensity by means of a thermopile.^{[57][58]}

In 1880, André-Prospere-Paul Crova published a diagram of the three-dimensional appearance of the graph of the strength of thermal radiation as a function of wavelength and temperature.^[59] He determined the spectral variable by use of prisms. He analyzed the surface through what he called "isothermal" curves, sections for a single temperature, with a spectral variable on the abscissa and a power variable on the ordinate. He put smooth curves through his experimental data points. They had one peak at a spectral value characteristic for the temperature, and fell either side of it towards the horizontal axis.^{[60][61]} Such spectral sections are widely shown even today.

In a series of papers from 1881 to 1886, Langley reported measurements of the spectrum of heat radiation, using diffraction gratings and prisms, and the most sensitive detectors that he could make. He reported that there was a peak intensity that increased with temperature, that the shape of the spectrum was not symmetrical about the peak, that there was a strong fall-off of intensity when the wavelength was shorter than an approximate cut-off value for each temperature, that the approximate cut-off wavelength decreased with increasing temperature, and that the wavelength of the peak intensity decreased with temperature, so that the intensity increased strongly with temperature for short wavelengths that were longer than the approximate cut-off for the temperature.^[62]

Having read Langley, in 1888, Russian physicist V.A. Michelson published a consideration of the idea that the unknown Kirchhoff radiation function could be explained physically and stated mathematically in terms of "complete irregularity of the vibrations of ... atoms".^{[63][64]} At this time, Planck was not studying radiation closely, and believed in neither atoms nor statistical physics.^[65] Michelson produced a formula for the spectrum for temperature:

$$I_\lambda = B_1 \theta^{\frac{3}{2}} \exp\left(-\frac{c}{\lambda^2 \theta}\right) \lambda^{-6},$$

where I_λ denotes specific radiative intensity at wavelength λ and temperature θ , and where B_1 and C are empirical constants.

In 1898, Otto Lummer and Ferdinand Kurlbaum published an account of their cavity radiation source.^[66] Their design has been used largely unchanged for radiation measurements to the present day. It was a platinum box, divided by diaphragms, with its interior blackened with iron oxide. It was an important ingredient for the progressively improved measurements that led to the discovery of Planck's law.^[67] A version described in 1901 had its interior blackened with a mixture of chromium, nickel, and cobalt oxides.^[68]

The importance of the Lummer and Kurlbaum cavity radiation source was that it was an experimentally accessible source of black-body radiation, as distinct from radiation from a simply exposed incandescent solid body, which had been the nearest available experimental approximation to black-body radiation over a suitable range of temperatures. The simply exposed incandescent solid bodies, that had been used before, emitted radiation with departures from the black-body spectrum that made it impossible to find the true black-body spectrum from experiments.^{[69][70]}

Planck's views before the empirical facts led him to find his eventual law

Planck first turned his attention to the problem of black body radiation in 1897.^[71] Theoretical and empirical progress enabled Lummer and Pringsheim to write in 1899 that available experimental evidence was approximately consistent with the specific intensity law $C\lambda^{-5}e^{-c/\lambda T}$ where C and c denote empirically measurable constants, and where λ and T denote wavelength and temperature respectively.^{[72][73]} For theoretical reasons, Planck at that time accepted this formulation, which has an effective cut-off of short wavelengths.^{[74][75][76]}

Finding the empirical law

Max Planck produced his law on 19 October 1900^{[77][78]} as an improvement upon the Wien approximation, published in 1896 by Wilhelm Wien, which fit the experimental data at short wavelengths (high frequencies) but deviated from it at long wavelengths (low frequencies).^[36] In June 1900, based on heuristic theoretical considerations, Rayleigh had suggested a formula^[79] that he proposed might be checked experimentally. The suggestion was that the Stewart–Kirchhoff universal function might be of the form $c_1 T \lambda^{-4} \exp(-\frac{c_2}{\lambda T})$. This was not the celebrated Rayleigh–Jeans formula $8\pi k_B T \lambda^{-4}$, which did not emerge until 1905,^[33] though it did reduce to the latter for long wavelengths, which are the relevant ones here. According to Klein,^[71] one may speculate that it is likely that Planck had seen this suggestion though he did not mention it in his papers of 1900 and 1901. Planck would have been aware of various other proposed formulas which had been offered.^{[55][80]} On 7 October 1900, Rubens told Planck that in the complementary domain (long wavelength, low frequency), and only there, Rayleigh's 1900 formula fitted the observed data well.^[80]

For long wavelengths, Rayleigh's 1900 heuristic formula approximately meant that energy was proportional to temperature,

$U_\lambda = \text{const. } T$.^{[71][80][81]} It is known that $\frac{dS}{dU_\lambda} = \frac{1}{T}$ and this leads to $\frac{dS}{dU_\lambda} = \frac{\text{const.}}{U_\lambda}$ and thence to $\frac{d^2S}{dU_\lambda^2} = -\frac{\text{const.}}{U_\lambda^2}$ for long wavelengths.

But for short wavelengths, the Wien formula leads to $\frac{1}{T} = -\text{const.} \ln U_\lambda + \text{const.}$ and thence to $\frac{d^2S}{dU_\lambda^2} = -\frac{\text{const.}}{U_\lambda}$ for short wavelengths.

Planck perhaps patched together these two heuristic formulas, for long and for short wavelengths,^{[80][82]} to produce a formula

$$\frac{d^2S}{dU_\lambda^2} = \frac{\alpha}{U_\lambda(\beta + U_\lambda)}. \quad [77]$$

This led Planck to the formula

$$B_\lambda(T) = \frac{C\lambda^{-5}}{e^{\frac{c}{\lambda T}} - 1},$$

where Planck used the symbols C and c to denote empirical fitting constants.

Planck sent this result to Rubens, who compared it with his and Kurlbaum's observational data and found that it fitted for all wavelengths remarkably well. On 19 October 1900, Rubens and Kurlbaum briefly reported the fit to the data,^[83] and Planck added a short presentation to give a theoretical sketch to account for his formula.^[77] Within a week, Rubens and Kurlbaum gave a fuller report of their measurements confirming Planck's law. Their technique for spectral resolution of the longer wavelength radiation was called the residual ray method. The rays were repeatedly reflected from polished crystal surfaces, and the rays that made it all the way through the process were 'residual', and were of wavelengths preferentially reflected by crystals of suitably specific materials.^{[84][85][86]}

Trying to find a physical explanation of the law

Once Planck had discovered the empirically fitting function, he constructed a physical derivation of this law. His thinking revolved around entropy rather than being directly about temperature. Planck considered a cavity with perfectly reflective walls; the cavity contained finitely many hypothetical well separated and recognizable but identically constituted, of definite magnitude, resonant oscillatory bodies, several such oscillators at each of finitely many characteristic frequencies. The hypothetical oscillators were for Planck purely imaginary theoretical

investigative probes, and he said of them that such oscillators do not need to "really exist somewhere in nature, provided their existence and their properties are consistent with the laws of thermodynamics and electrodynamics."^[87] Planck did not attribute any definite physical significance to his hypothesis of resonant oscillators, but rather proposed it as a mathematical device that enabled him to derive a single expression for the black body spectrum that matched the empirical data at all wavelengths.^[88] He tentatively mentioned the possible connection of such oscillators with atoms. In a sense, the oscillators corresponded to Planck's speck of carbon; the size of the speck could be small regardless of the size of the cavity, provided the speck effectively transduced energy between radiative wavelength modes.^[80]

Partly following a heuristic method of calculation pioneered by Boltzmann for gas molecules, Planck considered the possible ways of distributing electromagnetic energy over the different modes of his hypothetical charged material oscillators. This acceptance of the probabilistic approach, following Boltzmann, for Planck was a radical change from his former position, which till then had deliberately opposed such thinking proposed by Boltzmann.^[89] Heuristically, Boltzmann had distributed the energy in arbitrary merely mathematical quanta ϵ , which he had proceeded to make tend to zero in magnitude, because the finite magnitude ϵ had served only to allow definite counting for the sake of mathematical calculation of probabilities, and had no physical significance. Referring to a new universal constant of nature, h ,^[90] Planck supposed that, in the several oscillators of each of the finitely many characteristic frequencies, the total energy was distributed to each in an integer multiple of a definite physical unit of energy, ϵ , not arbitrary as in Boltzmann's method, but now for Planck, in a new departure, characteristic of the respective characteristic frequency.^{[78][91][92][93]} His new universal constant of nature, h , is now known as Planck's constant.

Planck explained further^[78] that the respective definite unit, ϵ , of energy should be proportional to the respective characteristic oscillation frequency ν of the hypothetical oscillator, and in 1901 he expressed this with the constant of proportionality h :^{[94][95]}

$$\epsilon = h\nu.$$

Planck did not propose that light propagating in free space is quantized.^{[96][97][98]} The idea of quantization of the free electromagnetic field was developed later, and eventually incorporated into what we now know as quantum field theory.^[99]

In 1906 Planck acknowledged that his imaginary resonators, having linear dynamics, did not provide a physical explanation for energy transduction between frequencies.^{[100][101]} Present-day physics explains the transduction between frequencies in the presence of atoms by their quantum excitability, following Einstein. Planck believed that in a cavity with perfectly reflecting walls and with no matter present, the electromagnetic field cannot exchange energy between frequency components.^[102] This is because of the linearity of Maxwell's equations.^[103] Present-day quantum field theory predicts that, in the absence of matter, the electromagnetic field obeys nonlinear equations and in that sense does self-interact.^{[104][105]} Such interaction in the absence of matter has not yet been directly measured because it would require very high intensities and very sensitive and low-noise detectors, which are still in the process of being constructed.^{[104][106]} Planck believed that a field with no interactions neither obeys nor violates the classical principle of equipartition of energy,^{[107][108]} and instead remains exactly as it was when introduced, rather than evolving into a black body field.^[109] Thus, the linearity of his mechanical assumptions precluded Planck from having a mechanical explanation of the maximization of the entropy of the thermodynamic equilibrium thermal radiation field. This is why he had to resort to Boltzmann's probabilistic arguments.^{[110][111]}

Planck's law may be regarded as fulfilling the prediction of Gustav Kirchhoff that his law of thermal radiation was of the highest importance. In his mature presentation of his own law, Planck offered a thorough and detailed theoretical proof for Kirchhoff's law,^[112] theoretical proof of which until then had been sometimes debated, partly because it was said to rely on unphysical theoretical objects, such as Kirchhoff's perfectly absorbing infinitely thin black surface.^[113]

Subsequent events

It was not till five years after Planck made his heuristic assumption of abstract elements of energy or of action that Albert Einstein conceived of really existing quanta of light in 1905^[114] as a revolutionary explanation of black-body radiation, of photoluminescence, of the photoelectric effect, and of the ionization of gases by ultraviolet light. In 1905, "Einstein believed that Planck's theory could not be made to agree with the idea of light quanta, a mistake he corrected in 1906."^[115] Contrary to Planck's beliefs of the time, Einstein proposed a model and formula whereby light was emitted, absorbed, and propagated in free space in energy quanta localized in points of space.^[114] As an introduction to his reasoning, Einstein recapitulated Planck's model of hypothetical resonant material electric oscillators as sources and sinks of radiation, but then he offered a new argument, disconnected from that model, but partly based on a thermodynamic argument of Wien, in which Planck's formula $\epsilon = h\nu$ played no role.^[116] Einstein gave the energy content of such quanta in the form $\frac{R\beta\nu}{N}$. Thus Einstein was contradicting the undulatory theory of light held by Planck. In 1910, criticizing a manuscript sent to him by Planck, knowing that Planck was a steady supporter of Einstein's theory of special relativity, Einstein wrote to Planck: "To me it seems absurd to have energy continuously distributed in space without assuming an aether."^[117]

According to Thomas Kuhn, it was not till 1908 that Planck more or less accepted part of Einstein's arguments for physical as distinct from abstract mathematical discreteness in thermal radiation physics. Still in 1908, considering Einstein's proposal of quantal propagation, Planck opined that such a revolutionary step was perhaps unnecessary.^[118] Until then, Planck had been consistent in thinking that discreteness of action quanta was to be found neither in his resonant oscillators nor in the propagation of thermal radiation. Kuhn wrote that, in Planck's earlier papers and in his 1906 monograph,^[119] there is no "mention of discontinuity, [nor] of talk of a restriction on oscillator energy, [nor of] any formula like $U = nh\nu$."^[120] Kuhn pointed out that his study of Planck's papers of 1900 and 1901, and of his monograph of

1906,^[119] had led him to "heretical" conclusions, contrary to the widespread assumptions of others who saw Planck's writing only from the perspective of later, anachronistic, viewpoints.^{[121][122]} Kuhn's conclusions, finding a period till 1908, when Planck consistently held his 'first theory', have been accepted by other historians.^{[123][124]}

In the second edition of his monograph, in 1912, Planck sustained his dissent from Einstein's proposal of light quanta. He proposed in some detail that absorption of light by his virtual material resonators might be continuous, occurring at a constant rate in equilibrium, as distinct from quantal absorption. Only emission was quantal.^{[103][125]} This has at times been called Planck's "second theory".^[126]

It was not till 1919 that Planck in the third edition of his monograph more or less accepted his 'third theory', that both emission and absorption of light were quantal.^[127]

The colourful term "ultraviolet catastrophe" was given by Paul Ehrenfest in 1911 to the paradoxical result that the total energy in the cavity tends to infinity when the equipartition theorem of classical statistical mechanics is (mistakenly) applied to black body radiation.^{[128][129]} But this had not been part of Planck's thinking, because he had not tried to apply the doctrine of equipartition: when he made his discovery in 1900, he had not noticed any sort of "catastrophe".^{[74][75][76][71][130]} It was first noted by Lord Rayleigh in 1900,^{[79][131][132]} and then in 1901^[133] by Sir James Jeans; and later, in 1905, by Einstein when he wanted to support the idea that light propagates as discrete packets, later called 'photons', and by Rayleigh^[34] and by Jeans.^{[33][134][135][136]}

In 1913, Bohr gave another formula with a further different physical meaning to the quantity $h\nu$.^{[29][30][31][137][138][139]} In contrast to Planck's and Einstein's formulas, Bohr's formula referred explicitly and categorically to energy levels of atoms. Bohr's formula was $W_{\tau_2} - W_{\tau_1} = h\nu$ where W_{τ_2} and W_{τ_1} denote the energy levels of quantum states of an atom, with quantum numbers τ_2 and τ_1 . The symbol ν denotes the frequency of a quantum of radiation that can be emitted or absorbed as the atom passes between those two quantum states. In contrast to Planck's model, the frequency ν has no immediate relation to frequencies that might describe those quantum states themselves.

Later, in 1924, Satyendra Nath Bose developed the theory of the statistical mechanics of photons, which allowed a theoretical derivation of Planck's law. The actual word 'photon' was invented still later, by G.N. Lewis in 1926,^[140] who mistakenly believed that photons were conserved, contrary to Bose–Einstein statistics; nevertheless the word 'photon' was adopted to express the Einstein postulate of the packet nature of light propagation. In an electromagnetic field isolated in a vacuum in a vessel with perfectly reflective walls, such as was considered by Planck, indeed the photons would be conserved according to Einstein's 1905 model, but Lewis was referring to a field of photons considered as a system closed with respect to ponderable matter but open to exchange of electromagnetic energy with a surrounding system of ponderable matter, and he mistakenly imagined that still the photons were conserved, being stored inside atoms.

Ultimately, Planck's law of black-body radiation contributed to Einstein's concept of quanta of light carrying linear momentum,^{[29][114]} which became the fundamental basis for the development of quantum mechanics.

The above-mentioned linearity of Planck's mechanical assumptions, not allowing for energetic interactions between frequency components, was superseded in 1925 by Heisenberg's original quantum mechanics. In his paper submitted on 29 July 1925, Heisenberg's theory accounted for Bohr's above-mentioned formula of 1913. It admitted non-linear oscillators as models of atomic quantum states, allowing energetic interaction between their own multiple internal discrete Fourier frequency components, on the occasions of emission or absorption of quanta of radiation. The frequency of a quantum of radiation was that of a definite coupling between internal atomic meta-stable oscillatory quantum states.^{[141][142]} At that time, Heisenberg knew nothing of matrix algebra, but Max Born read the manuscript of Heisenberg's paper and recognized the matrix character of Heisenberg's theory. Then Born and Jordan published an explicitly matrix theory of quantum mechanics, based on, but in form distinctly different from, Heisenberg's original quantum mechanics; it is the Born and Jordan matrix theory that is today called matrix mechanics.^{[143][144][145]} Heisenberg's explanation of the Planck oscillators, as non-linear effects apparent as Fourier modes of transient processes of emission or absorption of radiation, showed why Planck's oscillators, viewed as enduring physical objects such as might be envisaged by classical physics, did not give an adequate explanation of the phenomena.

Nowadays, as a statement of the energy of a light quantum, often one finds the formula $E = \hbar\omega$, where $\hbar = \frac{h}{2\pi}$, and $\omega = 2\pi\nu$ denotes angular frequency,^{[146][147][148][149][150]} and less often the equivalent formula $E = h\nu$.^{[149][150][151][152][153]} This statement about a really existing and propagating light quantum, based on Einstein's, has a physical meaning different from that of Planck's above statement $\epsilon = h\nu$ about the abstract energy units to be distributed amongst his hypothetical resonant material oscillators.

An article by Helge Kragh published in *Physics World* gives an account of this history.^[93]

See also

- Emissivity
- Radiance
- Sakuma–Hattori equation
- Wien's displacement law

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External links

- Summary of Radiation
- Radiation of a Blackbody – interactive simulation to play with Planck's law
- Scienceworld entry on Planck's Law

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Categories: Statistical mechanics | Foundational quantum physics | Max Planck

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Planetary equilibrium temperature

From Wikipedia, the free encyclopedia

The **planetary equilibrium temperature** is a theoretical temperature that a planet would be at when considered simply as if it were a black body being heated only by its parent star. In this model, the presence or absence of an atmosphere (and therefore any greenhouse effect) is not considered, and one treats the theoretical black body temperature as if it came from an idealized surface of the planet.

Other authors use different names for this concept, such as **equivalent blackbody temperature** of a planet,^[1] or the **effective radiation emission temperature** of the planet.^[2] Similar concepts include the global mean temperature, global radiative equilibrium, and global-mean surface air temperature,^[3] which includes the effects of global warming.

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Calculation of semi-blackbody temperature

If the incident solar radiation ("insolation") on the planet at its orbital distance from the Sun is I_o , the amount of energy absorbed by the planet will depend on its albedo a and its cross-sectional area:

$$P_{in} = I_o (1 - a) \pi R_p^2$$

Note that the albedo would be zero ($a = 0$) for a blackbody. However, in planetology, more useful results are obtained by accounting for a measured or assumed planetary albedo $a > 0$.

The infrared power radiated by the planet as thermal radiation will depend on its emissivity and its surface area, according to the Stefan–Boltzmann equation:

$$P_{out} = \epsilon \sigma A T^4$$

where P_{out} is the radiated power, ϵ is the emissivity, σ the Stefan–Boltzmann constant, A the surface area, and T the absolute temperature. For a spherical planet, the surface area is $A = 4\pi R_p^2$.

The emissivity is typically assumed to be $\epsilon = 1$, as would be the case for a perfectly emitting blackbody. This is often a good assumption, as emissivities of natural surfaces tend to be in the range of 0.9 to 1, e.g.

$$\epsilon_{Earth(Planet)} = 0.96.$$

The equilibrium temperature is then calculated by setting $P_{in} = P_{out}$. Thus,

$$T_{eq} = \left(\frac{I_o (1 - a)}{4\sigma} \right)^{1/4}$$

Theoretical model

Consider a spherical star and a spherical planet. The star and the planet are considered to be perfect black bodies. The planet has an albedo and only absorbs a fraction of radiation, depending on its surface characteristics. The star emits radiation isotropically according to the Stefan–Boltzmann law which travels a distance equal to the orbital distance of the planet, D . The planet absorbs the radiation that isn't reflected by the albedo of the surface, and heats up. Since the planet is also a black body which emits radiation according to the Stefan–Boltzmann law, it will emit radiation and lose energy. Thermal equilibrium exists when the power supplied by the star is equal to the power emitted by the planet. The temperature at which this balance occurs is the planetary equilibrium temperature and is equal to:

$$T_{eq} = T_{\odot} (1 - a)^{1/4} \sqrt{\frac{R_{\odot}}{2D}}$$

Where T_{\odot} and R_{\odot} are the temperature and radius of the star.

The equilibrium temperature is neither an upper nor lower bound on actual temperatures on a planet. Because of the greenhouse effect, planets with atmospheres will have temperatures higher than the equilibrium temperature. For example, Venus has an equilibrium temperature of approximately 227 K, but a surface temperature of 740 K.^{[4][5]} The Moon has a black body temperature of 271 K,^[6] but can have temperatures of 373 K in the daytime and 100 K at night.^[7] This is due to the relatively slow rotation of the Moon compared to its size, so that the entire surface doesn't heat evenly. Orbiting bodies can also be heated by Tidal heating,^[8] Geothermal energy which is driven by radioactive decay in the core of the planet,^[9] or accretional heating.^[10]

Detailed derivation of the planetary equilibrium temperature

The power absorbed by the planet from the star is equal to the power emitted by the planet: $P_{in} = P_{out}$

The power input to the planet is equal to the solar irradiance (*i.e.* power received per unit area) of the star at the distance of the planet, I_o , times the fraction absorbed by the planet (1 minus the albedo), times the area of the planet illuminated by the star: $P_{in} = I_o (1 - a) \pi R_p^2$

I_o , the solar intensity at the distance of the planet from the sun, is equal to the luminosity (*i.e.* total power emitted) of the star divided by the area of the sphere that all of the star's radiation is cast on at the distance of the planet. This yields: $P_{in} = L_{\odot} (1 - a) \left(\frac{\pi R_p^2}{4\pi D^2} \right)$ ^[5]

Any incoming power to a black body is radiated as heat according to the Stefan–Boltzmann law $P = \epsilon \sigma A T^4$.

(The emissivity ϵ is usually expected to be very close to 1, and thus is often left out). Multiplying by the area, the power emitted by the planet is: $P_{out} = (\epsilon \sigma T_{eq}^4) (4\pi R_p^2)$

Setting these equal:

$$T_{eq} = \left(\frac{L_{\odot} (1 - a)}{16\epsilon\sigma\pi D^2} \right)^{1/4}$$

The luminosity of the star is equal to the Stefan-Boltzmann constant, times the area of the star, times the fourth power of the temperature of the star: $L_{\odot} = (\sigma T_{\odot}^4) (4\pi R_{\odot}^2)$

Inserting this into the previous equation, it can be shown that:

$$T_{eq} = T_{\odot} \left(\frac{(1 - a)}{\epsilon} \right)^{1/4} \sqrt{\frac{R_{\odot}}{2D}}$$

By assuming that the emissivity ϵ is equal to 1, this reproduces the equation in the previous section. It is interesting to note that the equilibrium temperature does not depend on the size of the planet, because both the incoming radiation and outgoing radiation depend on the area of the planet.

Calculation for extrasolar planets

For extrasolar planets the temperature of the star can be calculated from the color of the star using Planck's law. The calculated temperature of the star can be used with the Hertzsprung–Russell diagram to determine the absolute magnitude of the star, which can then be used with observational data to determine the distance to the star and finally the size of the star. Orbital simulations are used to determine what orbital parameters (including orbital distance) produce the observations seen by astronomers.^[11] Astronomers use a hypothesized albedo^[12] and can then estimate the equilibrium temperature.

See also

- Effective temperature
- Thermal equilibrium

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External links

- Equilibrium Temperature at the Laboratory for Atmospheric and Space Physics, University of Colorado
- Energy balance: the simplest climate model
- HEC: Exoplanets Calculator Features a user friendly calculator to calculate the Planet Equilibrium Temperature.

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Rayleigh scattering

From Wikipedia, the free encyclopedia

Rayleigh scattering (pronounced /ˈreɪli/ *RAY-lee*), named after the British physicist Lord Rayleigh (John William Strutt),^[1] is the (dominantly) elastic scattering of light or other electromagnetic radiation by particles much smaller than the wavelength of the radiation. Rayleigh scattering does not change the state of material and is, hence, a parametric process. The particles may be individual atoms or molecules. It can occur when light travels through transparent solids and liquids, but is most prominently seen in gases. Rayleigh scattering results from the electric polarizability of the particles. The oscillating electric field of a light wave acts on the charges within a particle, causing them to move at the same frequency. The particle therefore becomes a small radiating dipole whose radiation we see as scattered light.



Rayleigh scattering causes the blue hue of the daytime sky and the reddening of the sun at sunset.

Rayleigh scattering of sunlight in the atmosphere causes diffuse sky radiation, which is the reason for the blue color of the sky and the yellow tone of the sun itself.

The amount of scattering is inversely proportional to the fourth power of the wavelength.

Rayleigh scattering of molecular nitrogen and oxygen in the atmosphere includes elastic scattering as well as the inelastic contribution from rotational Raman scattering in air, since the changes in wavenumber of the scattered photon are typically smaller than 50 cm^{-1} .^[2] This can lead to changes in the rotational state of the molecules. Furthermore, the inelastic contribution has the same wavelengths dependency as the elastic part.

Scattering by particles similar to, or larger than, the wavelength of light is typically treated by the Mie theory, the discrete dipole approximation and other computational techniques. Rayleigh scattering applies to particles that are small with respect to wavelengths of light, and that are optically "soft" (i.e. with a refractive index close to 1). On the other hand, anomalous diffraction theory applies to optically soft but larger particles.

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Small size parameter approximation

The size of a scattering particle is often parameterized by the ratio

$$x = \frac{2\pi r}{\lambda}$$

where r is its characteristic length (radius) and λ is the wavelength of the light. The amplitude of light scattered from within any transparent dielectric is proportional to the inverse square of its wavelength and to the volume of material, that is to the cube of its characteristic length. The wavelength dependence is characteristic of dipole scattering^[3] and the volume dependence will apply to any scattering mechanism. Objects with $x \gg 1$ act as geometric shapes, scattering light according to their projected area. At the intermediate $x \approx 1$ of Mie scattering, interference effects develop through phase variations over the object's surface. Rayleigh scattering applies to the case when the scattering particle is very small ($x \ll 1$, with a particle size $< 1/10$ wavelength^[4]) and the whole surface re-radiates with the same phase. Because the particles are randomly positioned, the scattered light arrives at a particular point with a random collection of phases; it is incoherent and the resulting intensity is just the sum of the squares of the amplitudes from each particle and therefore proportional to the inverse fourth power of the wavelength and the sixth power of its size.^{[3][5]} In detail, the intensity I of light scattered by any one of the small spheres of diameter d and refractive index n from a beam of unpolarized light of wavelength λ and intensity I_0 is given by

$$I = I_0 \frac{1 + \cos^2 \theta}{2R^2} \left(\frac{2\pi}{\lambda} \right)^4 \left(\frac{n^2 - 1}{n^2 + 2} \right)^2 \left(\frac{d}{2} \right)^6$$

[6]

where R is the distance to the particle and θ is the scattering angle. Averaging this over all angles gives the Rayleigh scattering cross-section^[7]

$$\sigma_s = \frac{2\pi^5}{3} \frac{d^6}{\lambda^4} \left(\frac{n^2 - 1}{n^2 + 2} \right)^2$$

[8]

The fraction of light scattered by a group of scattering particles is the number of particles per unit volume N times the cross-section. For example, the major constituent of the atmosphere, nitrogen, has a Rayleigh cross section of $5.1 \times 10^{-31} \text{ m}^2$ at a wavelength of 532 nm (green light).^[9] This means that at atmospheric pressure, where there are about 2×10^{25} molecules per cubic meter, about a fraction 10^{-5} of the light will be scattered for every meter of travel.

The strong wavelength dependence of the scattering ($\sim \lambda^{-4}$) means that shorter (blue) wavelengths are scattered more strongly than longer (red) wavelengths.

From molecules

The expression above can also be written in terms of individual molecules by expressing the dependence on refractive index in terms of the molecular polarizability α , proportional to the dipole moment induced by the electric field of the light. In this case, the Rayleigh scattering intensity for a single particle is given in CGS-units by^[10]

$$I = I_0 \frac{8\pi^4 \alpha^2}{\lambda^4 R^2} (1 + \cos^2 \theta).$$

Cause of the blue color of the sky

The strong wavelength dependence of the scattering ($\sim\lambda^{-4}$) means that shorter (blue) wavelengths are scattered more strongly than longer (red) wavelengths. This results in the indirect blue light coming from all regions of the sky. Rayleigh scattering is a good approximation of the manner in which light scattering occurs within various media for which scattering particles have a small size (parameter).

A portion of the beam of light coming from the sun scatters off molecules of gas and other small particles in the atmosphere. Here, Rayleigh scattering primarily occurs through sunlight's interaction with randomly located air molecules. It is this scattered light that gives the surrounding sky its brightness and its color. As previously stated, Rayleigh scattering is inversely proportional to the fourth power of wavelength, so that shorter wavelength violet and blue light will scatter more than the longer wavelengths (yellow and especially red light). However, the Sun, like any star, has its own spectrum and so I_0 in the scattering formula above is not constant but falls away in the violet. In addition the oxygen in the Earth's atmosphere absorbs wavelengths at the edge of the ultra-violet region of the spectrum. The resulting color, which appears like a pale blue, actually is a mixture of all the scattered colors, mainly blue and green. Conversely, glancing toward the sun, the colors that were not scattered away — the longer wavelengths such as red and yellow light — are directly visible, giving the sun itself a slightly yellowish hue. Viewed from space, however, the sky is black and the sun is white.

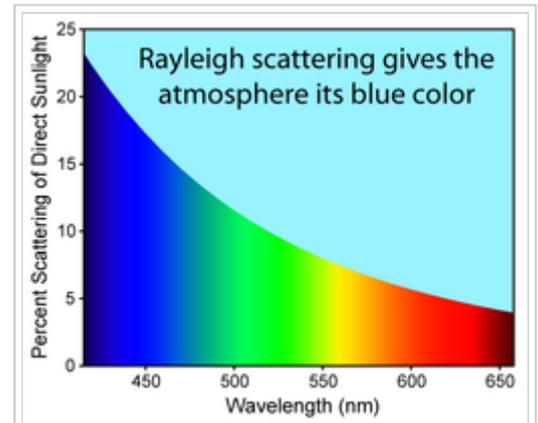


Figure showing the greater proportion of blue light scattered by the atmosphere relative to red light.



Scattered blue light is polarized. The picture on the right is shot through a polarizing filter: the polarizer transmits light that is linearly polarized in a specific direction.

The reddening of the sun is intensified when it is near the horizon because the light being received directly from it must pass through more of the atmosphere. The effect is further increased because the sunlight must pass through a greater proportion of the atmosphere nearer the earth's surface, where it is denser. This removes a significant proportion of the shorter wavelength (blue) and medium wavelength (green) light from the direct path to the observer. The remaining unscattered light is therefore mostly of longer wavelengths and appears redder.

Some of the scattering can also be from sulfate particles. For years after large Plinian eruptions, the blue cast of the sky is notably brightened by the persistent sulfate load of the stratospheric gases. Some works of the artist J. M. W. Turner may owe their vivid red colours to the eruption of Mount Tambora in his lifetime.^[11]

In locations with little light pollution, the moonlit night sky is also blue, because moonlight is reflected sunlight, with a slightly lower color temperature due to the brownish color of the moon. The moonlit sky is not perceived as blue, however, because at low light levels human vision comes mainly from rod cells that do not produce any color perception (Purkinje effect).

In optical fibers

Rayleigh scattering is an important component of the scattering of optical signals in optical fibers. Silica fibers are glasses, disordered materials with microscopic variations of density and refractive index. These give rise to energy losses due to the scattered light, with the following coefficient:^[12]

$$\alpha_{\text{scat}} = \frac{8\pi^3}{3\lambda^4} n^8 p^2 k T_f \beta$$

where n is the refraction index, p is the photoelastic coefficient of the glass, k is the Boltzmann constant, and β is the isothermal compressibility. T_f is a *fictive temperature*, representing the temperature at which the density fluctuations are "frozen" in the material.

In porous materials

λ^{-4} Rayleigh-type scattering can also be exhibited by porous materials. An example is the strong optical scattering by nanoporous materials.^[14] The strong contrast in refractive index between pores and solid parts of sintered alumina results in very strong scattering, with light completely changing direction each 5 micrometers on average. The λ^{-4} -type scattering is caused by the nanoporous structure (a narrow pore size distribution around ~70 nm) obtained by sintering monodisperse alumina powder.



Rayleigh scattering in opalescent glass: it appears blue from the side, but orange light shines through.^[13]

See also

- Rayleigh sky model
- Ricean fading
- Optical phenomenon
- Dynamic light scattering
- Tyndall effect
- Critical opalescence
- Marian Smoluchowski
- Rayleigh Criterion
- Aerial perspective
- Parametric Process
- Bragg's law

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- Pedro Lilienfeld, "A Blue Sky History" (2004). *Optics and Photonics News* Vol. 15, Issue 6, pp. 32–39. doi:10.1364/OPN.15.6.000032 Gives a brief history of theories of why the sky is blue leading up to Rayleigh's discovery, and a brief description of Rayleigh scattering.

External links

- HyperPhysics description of Rayleigh scattering
- Full physical explanation of sky color, in simple terms

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Categories: Scattering, absorption and radiative transfer (optics) | Atmospheric optical phenomena | Visibility | Light

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Rayleigh–Bénard convection

From Wikipedia, the free encyclopedia

Rayleigh–Bénard convection is a type of natural convection, occurring in a plane horizontal layer of fluid heated from below, in which the fluid develops a regular pattern of convection cells known as **Bénard cells**.

Rayleigh–Bénard convection is one of the most commonly studied convection phenomena because of its analytical and experimental accessibility.^[1] The convection patterns are the most carefully examined example of self-organizing nonlinear systems.^[1] ^[2]

Buoyancy, and hence gravity, is responsible for the appearance of convection cells. The initial movement is the upwelling of lesser density fluid from the heated bottom layer.^[3] This upwelling spontaneously organizes into a regular pattern of cells.



Play media

Bénard cells.

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Physical processes

The features of Rayleigh–Bénard convection can be obtained by a simple experiment first conducted by Henri Bénard, a French physicist, in 1900.

Development of convection

The experimental set-up uses a layer of liquid, e.g. water, between two parallel planes. The height of the layer is small compared to the horizontal dimension. At first, the temperature of the bottom plane is the same as the top plane. The liquid will then tend towards an equilibrium, where its temperature is the same as its surroundings. (Once there, the liquid is perfectly uniform: to an observer it would appear the same from any position. This equilibrium is also asymptotically stable: after a local, temporary perturbation of the outside temperature, it will go back to its uniform state, in line with the second law of thermodynamics).

Then, the temperature of the bottom plane is increased slightly yielding a flow of thermal energy conducted through the liquid. The system will begin to have a structure of thermal conductivity: the temperature, and the density and pressure with it, will vary linearly between the bottom and top plane. A uniform linear gradient of temperature will be established. (This system may be modelled by statistical mechanics).

Once conduction is established, the microscopic random movement *spontaneously* becomes ordered on a macroscopic level, forming Bénard convection cells, with a characteristic correlation length.

Convection features

The rotation of the cells is stable and will alternate from clock-wise to counter-clockwise horizontally; this is an example of spontaneous symmetry breaking. Bénard cells are metastable. This means that a small perturbation will not be able to change the rotation of the cells, but a larger one could affect the rotation; they exhibit a form of hysteresis.

Moreover, the deterministic law at the microscopic level produces a non-deterministic arrangement of the cells: if the experiment is repeated, a particular position in the experiment will be in a clockwise cell in some cases, and a counter-clockwise cell in others. Microscopic perturbations of the initial conditions are enough to produce a non-deterministic macroscopic effect. That is, in principle, there is no way to calculate the macroscopic effect of a microscopic perturbation. This inability to predict long-range conditions and sensitivity to initial-conditions are characteristics of chaotic or complex systems (i.e., the butterfly effect).

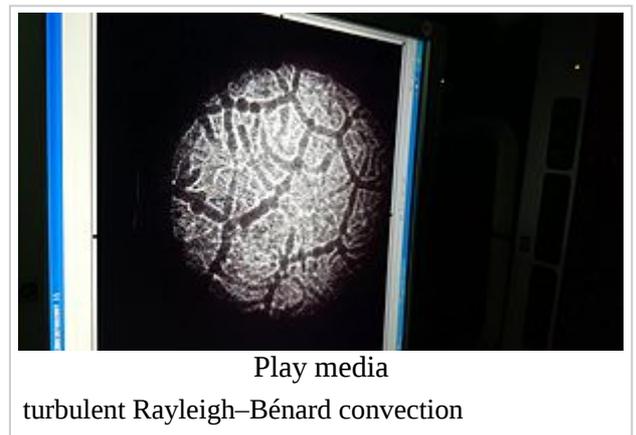
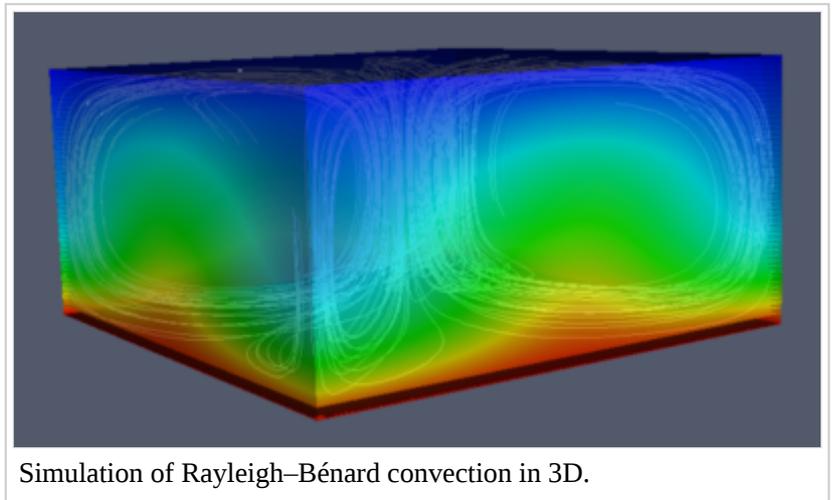
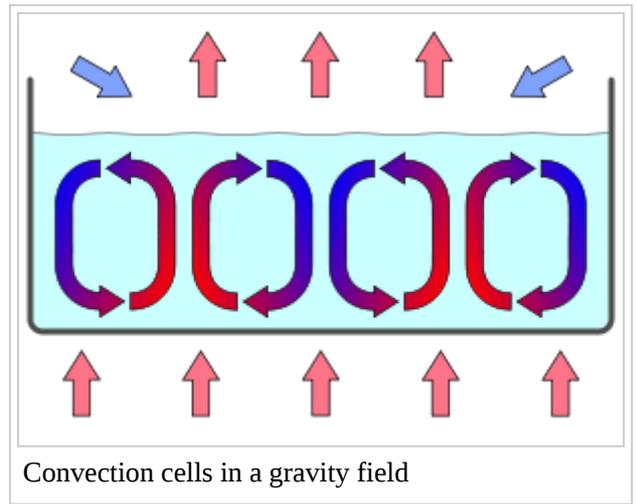
If the temperature of the bottom plane was to be further increased, the structure would become more complex in space and time; the turbulent flow would become chaotic.

Convective Bénard cells tend to approximate regular right hexagonal prisms, particularly in the absence of turbulence,^{[4][5][6]} although certain experimental conditions can result in the formation of regular right square prisms^[7] or spirals.^[8]

The convective Bénard cells are not unique and will usually appear only in the surface tension driven convection. In general the solutions to the Rayleigh and Pearson^[9] analysis (linear theory) assuming an infinite horizontal layer gives rise to degeneracy meaning that many patterns may be obtained by the system. Assuming uniform temperature at the top and bottom plates, when a realistic system is used (a layer with horizontal boundaries) the shape of the boundaries will mandate the pattern. More often than not the convection will appear as rolls or a superposition of them.

The Rayleigh–Bénard Instability

Since there is a density gradient between the top and the bottom plate, gravity acts trying to pull the cooler, denser liquid from the top to the bottom. This gravitational force is opposed by the viscous damping force in the fluid. The balance of these two forces is expressed by a non-dimensional parameter called the Rayleigh



number. The Rayleigh number is defined as:

$$\mathbf{Ra}_L = \frac{g\beta}{\nu\alpha}(T_b - T_u)L^3$$

where

T_u is the temperature of the top plate
 T_b is the temperature of the bottom plate
 L is the height of the container
 g is the acceleration due to gravity
 ν is the kinematic viscosity
 α is the Thermal diffusivity
 β is the Thermal expansion coefficient.

As the Rayleigh number increases, the gravitational forces become more dominant. At a critical Rayleigh number of 1708,^[2] instability sets in and convection cells appear.

The critical Rayleigh number can be obtained analytically for a number of different boundary conditions by doing a perturbation analysis on the linearized equations in the stable state.^[10] The simplest case is that of two free boundaries, which Lord Rayleigh solved in 1916,^[11] obtaining $Ra = \frac{27}{4}\pi^4 \approx 657.51$.^[12] In the case of a rigid boundary at the bottom and a free boundary at the top (as in the case of a kettle without a lid), the critical Rayleigh number comes out as $Ra = 1,100.65$.^[13]

Effects of surface tension

In case of a free liquid surface in contact with air, buoyancy and surface tension effects will also play a role in how the convection patterns develop. Liquids flow from places of lower surface tension to places of higher surface tension. This is called the Marangoni effect. When applying heat from below, the temperature at the top layer will show temperature fluctuations. With increasing temperature, surface tension decreases. Thus a lateral flow of liquid at the surface will take place,^[14] from warmer areas to cooler areas. In order to preserve a horizontal (or nearly horizontal) liquid surface, cooler surface liquid will descend. This down-welling of cooler liquid contributes to the driving force of the convection cells. The specific case of temperature gradient-driven surface tension variations is known as thermo-capillary convection, or Bénard–Marangoni convection.

History and nomenclature

Rayleigh–Bénard convection, whose effects are due solely to a temperature gradient, was first successfully analyzed by Lord Rayleigh; Rayleigh assumed boundary conditions in which the vertical velocity component and temperature disturbance vanish at the top and bottom boundaries (perfect thermal conduction). Those assumptions resulted in the analysis losing any connection with Henri Bénard's experiment. This resulted in discrepancies between theoretical and experimental results until Pearson reworked the problem based on surface tension.^[9] This is what was originally observed by Bénard. Nonetheless in modern usage "Rayleigh–Bénard convection" refers to the effects due to temperature, whereas "Bénard–Marangoni convection" refers specifically to the effects of surface tension.^[1] Davis and Koschmieder have suggested that the convection should be rightfully called the "Pearson–Bénard convection".^[2]

Rayleigh–Bénard convection is also sometimes known as "Bénard–Rayleigh convection", "Bénard convection", or "Rayleigh convection".

See also

- Hydrodynamic stability
- Marangoni effect
- Natural convection

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- K. Daniels, B. Plapp, W. Pesch, O. Brausch, E. Bodenschatz: Undulation Chaos in inclined Layer Convection
- Karen E. Daniel, Oliver Brausch, Werner Pesch, Eberhard Bodenschatz: Competition and bistability of ordered undulations and undulation chaos in inclined layer convection (PDF; 608 kB)
- P. Subramanian, O. Brausch, E. Bodenschatz, K. Daniels, T. Schneider W. Pesch: Spatio-temporal Patterns in Inclined Layer Convection (PDF; 5,3 MB)



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Stefan–Boltzmann law

From Wikipedia, the free encyclopedia

The **Stefan–Boltzmann law** describes the power radiated from a black body in terms of its temperature. Specifically, the Stefan–Boltzmann law states that the total energy radiated per unit surface area of a black body across all wavelengths per unit time (also known as the black-body *radiant emittance* or *radiant exitance*), j^* , is directly proportional to the fourth power of the black body's thermodynamic temperature T :

$$j^* = \sigma T^4.$$

The constant of proportionality σ , called the Stefan–Boltzmann constant derives from other known constants of nature. The value of the constant is

$$\sigma = \frac{2\pi^5 k^4}{15c^2 h^3} = 5.670373 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4},$$

where k is the Boltzmann constant, h is Planck's constant, and c is the speed of light in a vacuum. Thus at 100 K the energy flux is 5.67 W/m², at 1000 K 56,700 W/m², etc. The radiance (watts per square metre per steradian) is given by

$$L = \frac{j^*}{\pi} = \frac{\sigma}{\pi} T^4.$$

A body that does not absorb all incident radiation (sometimes known as a grey body) emits less total energy than a black body and is characterized by an emissivity, $\epsilon < 1$:

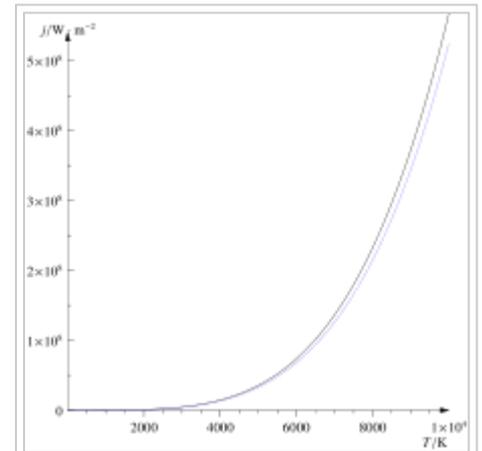
$$j^* = \epsilon \sigma T^4.$$

The irradiance j^* has dimensions of energy flux (energy per time per area), and the SI units of measure are joules per second per square metre, or equivalently, watts per square metre. The SI unit for absolute temperature T is the kelvin. ϵ is the emissivity of the grey body; if it is a perfect blackbody, $\epsilon = 1$. In the still more general (and realistic) case, the emissivity depends on the wavelength, $\epsilon = \epsilon(\lambda)$.

To find the total power radiated from an object, multiply by its surface area, A :

$$P = A j^* = A \epsilon \sigma T^4.$$

Wavelength- and subwavelength-scale particles,^[1] metamaterials,^[2] and other nanostructures are not subject to ray-optical limits and may be designed to exceed the Stefan–Boltzmann law.



Graph of a function of total emitted energy of a black body j^* proportional to its thermodynamic temperature T . In blue is a total energy according to the Wien approximation,

$$j_W^* = j^* / \zeta(4) \approx 0.924 \sigma T^4$$

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 - 2.3 Effective Temperature of the Earth
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- 5 Notes
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History

The law was deduced by Josef Stefan (1835–1893) in 1879 on the basis of experimental measurements made by John Tyndall and was derived from theoretical considerations, using thermodynamics, by Ludwig Boltzmann (1844–1906) in 1884. Boltzmann considered a certain ideal heat engine with light as a working matter instead of a gas. The law is highly accurate only for ideal black objects, the perfect radiators, called black bodies; it works as a good approximation for most "grey" bodies. Stefan published this law in the article *Über die Beziehung zwischen der Wärmestrahlung und der Temperatur* (*On the relationship between thermal radiation and temperature*) in the *Bulletins from the sessions* of the Vienna Academy of Sciences.

Examples

Temperature of the Sun

With his law Stefan also determined the temperature of the Sun's surface. He learned from the data of Charles Soret (1854–1904) that the energy flux density from the Sun is 29 times greater than the energy flux density of a certain warmed metal lamella (a thin plate). A round lamella was placed at such a distance from the measuring device that it would be seen at the same angle as the Sun. Soret estimated the temperature of the lamella to be approximately 1900 °C to 2000 °C. Stefan surmised that $\frac{1}{3}$ of the energy flux from the Sun is absorbed by the Earth's atmosphere, so he took for the correct Sun's energy flux a value $\frac{3}{2}$ times greater than Soret's value, namely $29 \times \frac{3}{2} = 43.5$.

Precise measurements of atmospheric absorption were not made until 1888 and 1904. The temperature Stefan obtained was a median value of previous ones, 1950 °C and the absolute thermodynamic one 2200 K. As $2.57^4 = 43.5$, it follows from the law that the temperature of the Sun is 2.57 times greater than the temperature of the lamella, so Stefan got a value of 5430 °C or 5700 K (the modern value is 5778 K^[3]). This was the first sensible value for the temperature of the Sun. Before this, values ranging from as low as 1800 °C to as high as 13,000,000 °C were claimed. The lower value of 1800 °C was determined by Claude Servais Mathias Pouillet (1790–1868) in 1838 using the Dulong-Petit law. Pouillet also took just half the value of the Sun's correct energy flux.

Temperature of stars

The temperature of stars other than the Sun can be approximated using a similar means by treating the emitted energy as a black body radiation.^[4] So:

$$L = 4\pi R^2 \sigma T_e^4$$

where **L** is the luminosity, **σ** is the Stefan–Boltzmann constant, **R** is the stellar radius and **T** is the effective temperature. This same formula can be used to compute the approximate radius of a main sequence star relative to the sun:

$$\frac{R}{R_\odot} \approx \left(\frac{T_\odot}{T} \right)^2 \cdot \sqrt{\frac{L}{L_\odot}}$$

where R_{\odot} , is the solar radius, L_{\odot} is the solar luminosity, and so forth.

With the Stefan–Boltzmann law, astronomers can easily infer the radii of stars. The law is also met in the thermodynamics of black holes in so-called Hawking radiation.

Effective Temperature of the Earth

Similarly we can calculate the effective temperature of the Earth T_{\oplus} by equating the energy received from the Sun and the energy radiated by the Earth, under the black-body approximation. The luminosity of the Sun, L_{\odot} , is given by:

$$L_{\odot} = 4\pi R_{\odot}^2 \sigma T_{\odot}^4$$

At Earth, this energy is passing through a sphere with a radius of a_0 , the distance between the Earth and the Sun, and the irradiance (received power per unit area) is given by

$$E_{\oplus} = \frac{L_{\odot}}{4\pi a_0^2}$$

The Earth has a radius of R_{\oplus} , and therefore has a cross-section of πR_{\oplus}^2 . The radiant flux (i.e. solar power) absorbed by the Earth is thus given by:

$$\Phi_{\text{abs}} = \pi R_{\oplus}^2 \times E_{\oplus} :$$

Assuming the exchange is in a steady state, the flux emitted by Earth must equal the flux absorbed, and so:

$$\begin{aligned} 4\pi R_{\oplus}^2 \sigma T_{\oplus}^4 &= \pi R_{\oplus}^2 \times E_{\oplus} \\ &= \pi R_{\oplus}^2 \times \frac{4\pi R_{\odot}^2 \sigma T_{\odot}^4}{4\pi a_0^2} \end{aligned}$$

T_{\oplus} can then be found:

$$\begin{aligned} T_{\oplus}^4 &= \frac{R_{\odot}^2 T_{\odot}^4}{4a_0^2} \\ T_{\oplus} &= T_{\odot} \times \sqrt{\frac{R_{\odot}}{2a_0}} \\ &= 5780 \text{ K} \times \sqrt{\frac{696 \times 10^6 \text{ m}}{2 \times 149.598 \times 10^9 \text{ m}}} \\ &\approx 279 \text{ K} \end{aligned}$$

where T_{\odot} is the temperature of the Sun, R_{\odot} the radius of the Sun, and a_0 is the distance between the Earth and the Sun. This gives an effective temperature of 6 °C on the surface of the Earth, assuming that it perfectly absorbs all emission falling on it and has no atmosphere.

The Earth has an albedo of 0.3, meaning that 30% of the solar radiation that hits the planet gets scattered back into space without absorption. The effect of albedo on temperature can be approximated by assuming that the energy absorbed is multiplied by 0.7, but that the planet still radiates as a black body (the latter by definition of effective temperature, which is what we are calculating). This approximation reduces the temperature by a factor of $0.7^{1/4}$, giving 255 K (−18 °C).^{[5][6]}

However, long-wave radiation from the surface of the earth is partially absorbed and re-radiated back down by greenhouse gases, namely water vapor, carbon dioxide and methane.^{[7][8]} Since the emissivity with greenhouse effect (weighted more in the longer wavelengths where the Earth radiates) is reduced more than the absorptivity (weighted more in the shorter wavelengths of the Sun's radiation) is reduced, the equilibrium temperature is higher than the simple black-body calculation estimates. As a result, the Earth's actual average surface temperature is about 288 K (15 °C), which is higher than the 255 K effective temperature, and even higher than the 279 K temperature that a black body would have.

In the above discussion, we have assumed that the whole surface of the earth is at one temperature. Another interesting question is to ask what the temperature of a blackbody surface on the earth would be assuming that it reaches equilibrium with the sunlight falling on it. This of course depends on the angle of the sun on the surface and on how much air the sunlight has gone through. When the sun is at the zenith and the surface is horizontal, the irradiance can be as high as 1120 W/m².^[9] The Stefan–Boltzmann law then gives a temperature of

$$T = \left(\frac{1120 \text{ W/m}^2}{\sigma} \right)^{1/4} \approx 375\text{K}$$

or 102 °C. (Above the atmosphere, the result is even higher: 394 K.) We can think of the earth's surface as "trying" to reach equilibrium temperature during the day, but being cooled by the atmosphere, and "trying" to reach equilibrium with starlight and possibly moonlight at night, but being warmed by the atmosphere.

Origination

Thermodynamic derivation of the energy density

[10]

The fact that the energy density of the box containing radiation is proportional to T^4 can be derived using thermodynamics. It follows from the Maxwell stress tensor of classical electrodynamics that the radiation pressure p is related to the internal energy density u :

$$p = \frac{u}{3}.$$

From the fundamental thermodynamic relation

$$dU = TdS - pdV,$$

we obtain the following expression, after dividing by dV and fixing T :

$$\left(\frac{\partial U}{\partial V} \right)_T = T \left(\frac{\partial S}{\partial V} \right)_T - p = T \left(\frac{\partial p}{\partial T} \right)_V - p.$$

The last equality comes from the following Maxwell relation:

$$\left(\frac{\partial S}{\partial V} \right)_T = \left(\frac{\partial p}{\partial T} \right)_V.$$

From the definition of energy density it follows that

$$U = uV$$

where the energy density of radiation only depends on the temperature, therefore

$$\left(\frac{\partial U}{\partial V}\right)_T = u \left(\frac{\partial V}{\partial V}\right)_T = u.$$

Now, the equality

$$\left(\frac{\partial U}{\partial V}\right)_T = T \left(\frac{\partial p}{\partial T}\right)_V - p,$$

after substitution of $\left(\frac{\partial U}{\partial V}\right)_T$ and p for the corresponding expressions, can be written as

$$u = \frac{T}{3} \left(\frac{\partial u}{\partial T}\right)_V - \frac{u}{3}.$$

Since the partial derivative $\left(\frac{\partial u}{\partial T}\right)_V$ can be expressed as a relationship between only u and T (if one isolates it on one side of the equality), the partial derivative can be replaced by the ordinary derivative. After separating the differentials the equality becomes

$$\frac{du}{4u} = \frac{dT}{T},$$

which leads immediately to $u = AT^4$, with A as some constant of integration.

Derivation from Planck's law

The law can be derived by considering a small flat black body surface radiating out into a half-sphere. This derivation uses spherical coordinates, with φ as the zenith angle and θ as the azimuthal angle; and the small flat blackbody surface lies on the xy -plane, where $\varphi = \pi/2$.

The intensity of the light emitted from the blackbody surface is given by Planck's law :

$$I(\nu, T) = \frac{2h\nu^3}{c^2} \frac{1}{e^{\frac{h\nu}{kT}} - 1}.$$

where

- $I(\nu, T)$ is the amount of power per unit surface area per unit solid angle per unit frequency emitted at a frequency ν by a black body at temperature T .
- h is Planck's constant
- c is the speed of light, and
- k is Boltzmann's constant.

The quantity $I(\nu, T) A d\nu d\Omega$ is the power radiated by a surface of area A through a solid angle $d\Omega$ in the frequency range between ν and $\nu + d\nu$.

The Stefan–Boltzmann law gives the power emitted per unit area of the emitting body,

$$\frac{P}{A} = \int_0^\infty I(\nu, T) d\nu \int d\Omega$$

To derive the Stefan–Boltzmann law, we must integrate Ω over the half-sphere and integrate ν from 0 to ∞ . Furthermore, because black bodies are *Lambertian* (i.e. they obey Lambert's cosine law), the intensity observed along the sphere will be the actual intensity times the cosine of the zenith angle ϕ , and in spherical coordinates, $d\Omega = \sin(\phi) d\phi d\theta$.

$$\begin{aligned}\frac{P}{A} &= \int_0^\infty I(\nu, T) d\nu \int_0^{2\pi} d\theta \int_0^{\pi/2} \cos \phi \sin \phi d\phi \\ &= \pi \int_0^\infty I(\nu, T) d\nu\end{aligned}$$

Then we plug in for I :

$$\frac{P}{A} = \frac{2\pi h}{c^2} \int_0^\infty \frac{\nu^3}{e^{\frac{h\nu}{kT}} - 1} d\nu$$

To do this integral, do a substitution,

$$u = \frac{h\nu}{kT}$$

$$du = \frac{h}{kT} d\nu$$

which gives:

$$\frac{P}{A} = \frac{2\pi h}{c^2} \left(\frac{kT}{h}\right)^4 \int_0^\infty \frac{u^3}{e^u - 1} du.$$

The integral on the right is standard and goes by many names: it is a particular case of a Bose-Einstein integral, or the Riemann zeta function, $\zeta(4)$, or the Polylogarithm. The value of the integral is $\frac{\pi^4}{15}$, giving the result that, for a perfect blackbody surface:

$$j^* = \sigma T^4, \quad \sigma = \frac{2\pi^5 k^4}{15c^2 h^3} = \frac{\pi^2 k^4}{60\hbar^3 c^2}.$$

Finally, this proof started out only considering a small flat surface. However, any differentiable surface can be approximated by a bunch of small flat surfaces. So long as the geometry of the surface does not cause the blackbody to reabsorb its own radiation, the total energy radiated is just the sum of the energies radiated by each surface; and the total surface area is just the sum of the areas of each surface—so this law holds for all convex blackbodies, too, so long as the surface has the same temperature throughout. The law extends to radiation from non-convex bodies by using the fact that the convex hull of a black body radiates as though it were itself a black body.

See also

- Wien's displacement law
- Rayleigh–Jeans law
- Radiance
- Zero-dimensional models
- Black body

- Sakuma–Hattori equation

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Categories: Laws of thermodynamics | Power laws | Heat transfer

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Terrestrial biological carbon cycle

From Wikipedia, the free encyclopedia

Carbon is an essential part of life on Earth. About half the dry weight of most living organisms is carbon. It plays an important role in the structure, biochemistry, and nutrition of all living cells. Living biomass holds between 600 and 1,000 gigatons of carbon, most of which is wood, while some 1,200 gigatons of carbon are stored in the terrestrial biosphere as dead biomass.^[1]

Carbon is cycled through the terrestrial biosphere with varying speeds, depending on what form it is stored in and under which circumstances.^[2] It is exchanged most quickly with the atmosphere, although small amounts of carbon leave the terrestrial biosphere and enter the oceans as dissolved organic carbon (DOC).

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- 2 Carbon exchange between the terrestrial biosphere and other systems
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 - 2.3 Anthropogenic influences
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Movement of carbon in the terrestrial biosphere

Most carbon in the terrestrial biosphere is stored in forests: they hold 86% of the planet's terrestrial above-ground carbon and forest soils also hold 73% of the planet's soil carbon.^[3] Carbon stored inside plants can be transferred into other organisms during plant consumption. When animals eat plants, for example, the organic carbon stored in the plants is converted into other forms and utilized inside the animals. The same is true for bacteria and other heterotrophs. Dead plant material in or above soils remains there for some time before being respired by heterotrophs. Thus carbon is transferred in every step of the food chain from one organism to another.

Carbon exchange between the terrestrial biosphere and other systems

Atmosphere

Autotrophs, such as trees and other green plants, use photosynthesis to convert carbon dioxide during primary production, releasing oxygen in the process. This process occurs most quickly in ecosystems with high amounts of growth, such as in young forests. Because carbon is consumed in the process of autotrophic growth, more carbon is consumed in spring and summer during daytime than in winter and at night, when photosynthesis no longer takes place in most plants. Carbon storage in the biosphere is influenced by a number of processes on different time-scales: while carbon uptake through autotrophic respiration follows a diurnal and seasonal cycle, carbon can be stored in the terrestrial biosphere for up to several centuries, e.g. in wood or soil.

Most carbon leaves the terrestrial biosphere through respiration. When oxygen is present, aerobic respiration occurs, producing carbon dioxide. If oxygen is not present, e.g. as is the case in marshes or in animals' digestive tracts, anaerobic respiration can occur, which produces methane. About half of the gross primary production is respired by plants directly back into the atmosphere. Part of the net primary production, or the remaining carbon absorbed by the biosphere, is emitted back into the atmosphere through fires and heterotrophic

respiration. The rest is converted into soil organic carbon, which is released more slowly, or "inert" dissolved carbon, which can remain in the biosphere for an unknown period of time.^[2]

Geosphere

Carbon in the terrestrial biosphere enters the geosphere only through highly specialized processes. When anaerobic decomposition converts organic material into hydrocarbon rich materials and is then deposited as sediment, the carbon can enter the geosphere through tectonic processes and remain there for several million years. This process can lead to the creation of fossil fuels.

Anthropogenic influences

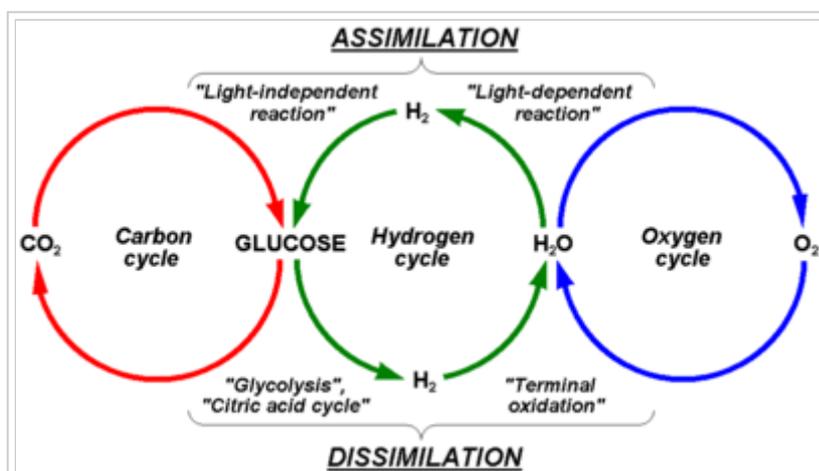
Human activity has large effects on the terrestrial biosphere, changing the way that it acts as a carbon reservoir. Anthropogenically caused fires release large amounts of carbon as CO₂ directly into the atmosphere. More significantly, however, humans modify land cover. Land cover change greatly decreases the amount of carbon uptake in the terrestrial biosphere. It modifies the local ecosystem, often replacing carbon-rich forest with agricultural or urban land use. This releases the carbon stored in the former land cover type and simultaneously decreases the biosphere's ability to absorb carbon from the atmosphere.

Indirectly, human-induced changes in the global climate cause widespread modifications to the terrestrial ecosystem's function in the carbon cycle. As local climates transition, locations that have long been conducive to one type of ecosystem can become more favorable for other ecosystem types. For example, warming in the Arctic has caused stress in North American boreal forests,^[4] thus decreasing primary production and carbon uptake, while the same warmer temperatures have led to increased shrub growth in the same areas,^[5] producing an opposite effect. Changes in weather patterns can also affect animals. For example, changed weather patterns may create favorable conditions for pine beetles, leading to large beetle outbreaks and forest destruction.^[6] Modified precipitation patterns can also lead to droughts or extreme precipitation events, causing additional stress for ecosystems and more erosion. Not only do such influences on the terrestrial ecosystem modify its carbon exchange with the atmosphere - they also can lead to increased outwashing of carbon into the oceans through the transport of organic material in rivers. These widespread changes in land cover also causes changes to the planetary albedo, inducing complex feedbacks in the Earth's planetary radiation budget.

Higher CO₂ levels in the atmosphere can cause photosynthesis to take place more efficiently, thus increasing plant growth and primary production. This could lead to the biosphere extracting more carbon dioxide from the atmosphere. How long this carbon would remain sequestered in the terrestrial biosphere before being rereleased into the atmosphere is unclear, however, and it is likely that other limiting factors (e.g. nitrogen availability, moisture, etc.) would prevent CO₂ fertilization from significantly increasing primary production.

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See also

Deep Carbon Observatory

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Thermal conduction

From Wikipedia, the free encyclopedia

Thermal conduction is the transfer of heat (internal energy) by microscopic collisions of particles and movement of electrons within a body. The microscopically colliding objects, that include molecules, atoms, and electrons, transfer disorganized microscopic kinetic and potential energy, jointly known as internal energy. Conduction takes place in all phases of matter, such as solids, liquids, gases and plasmas. The rate at which energy is conducted as heat between two bodies is a function of the temperature difference (temperature gradient) between the two bodies and the properties of the conductive medium through which the heat is transferred. Thermal conduction was originally called diffusion.

Heat spontaneously flows from a hotter to a colder body. For example, heat is conducted from the hotplate of an electric stove to the bottom of a saucepan in contact with it. In the absence of an external driving energy source to the contrary, within a body or between bodies, temperature differences decay over time, and thermal equilibrium is approached, temperature becoming more uniform.

In conduction, the heat flow is within and through the body itself. In contrast, in heat transfer by thermal radiation, the transfer is often between bodies, which may be separated spatially. Also possible is transfer of heat by a combination of conduction and thermal radiation. In convection, internal energy is carried between bodies by a moving material carrier. In solids, conduction is mediated by the combination of vibrations and collisions of molecules, of propagation and collisions of phonons, and of diffusion and collisions of free electrons. In gases and liquids, conduction is due to the collisions and diffusion of molecules during their random motion. Photons in this context do not collide with one another, and so heat transport by electromagnetic radiation is conceptually distinct from heat conduction by microscopic diffusion and collisions of material particles and phonons. But the distinction is often not easily observed, unless the material is semi-transparent.

In the engineering sciences, heat transfer includes the processes of thermal radiation, convection, and sometimes mass transfer. Usually, more than one of these processes occurs in a given situation. The conventional symbol for the material property, thermal conductivity, is *k*.

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Overview

On a microscopic scale, conduction occurs within a body considered as being stationary; this means that the kinetic and potential energies of the bulk motion of the body are separately accounted for. Internal energy diffuses as rapidly moving or vibrating atoms and molecules interact with neighboring particles, transferring some of their microscopic kinetic and potential energies, these quantities being defined relative to the bulk of the body considered as being stationary. Heat is transferred by conduction when adjacent atoms or molecules collide, or as several electrons move backwards and forwards from atom to atom in a disorganized way so as not to form a macroscopic electric current, or as phonons collide and scatter. Conduction is the most significant means of heat transfer within a solid or between solid objects in thermal contact. Conduction is greater in solids because the network of relatively close fixed spatial relationships between atoms helps to transfer energy between them by vibration.

Fluids (and especially gases) are less conductive. This is due to the large distance between atoms in a gas: fewer collisions between atoms means less conduction. At low densities, the conductivity of gases can be derived from kinetic theory using rigid, non-interacting spheres and refined using Chapman-Enskog kinetic theory of gases.^[1] The conductivity of gases increases with temperature. Conductivity increases with increasing pressure from vacuum up to a critical point that the density of the gas is such that molecules of the gas may be expected to collide with each other before they transfer heat from one surface to another. After this point, conductivity increases only slightly with increasing pressure and density.

Thermal contact conductance is the study of heat conduction between solid bodies in contact. A temperature drop is often observed at the interface between the two surfaces in contact. This phenomenon is said to be a result of a thermal contact resistance existing between the contacting surfaces. Interfacial thermal resistance is a measure of an interface's resistance to thermal flow. This thermal resistance differs from contact resistance, as it exists even at atomically perfect interfaces. Understanding the thermal resistance at the interface between two materials is of primary significance in the study of its thermal properties. Interfaces often contribute significantly to the observed properties of the materials.

The inter-molecular transfer of energy could be primarily by elastic impact, as in fluids, or by free electron diffusion, as in metals, or phonon vibration, as in insulators. In insulators, the heat flux is carried almost entirely by phonon vibrations.

Metals (e.g., copper, platinum, gold, etc.) are usually good conductors of thermal energy. This is due to the way that metals bond chemically: metallic bonds (as opposed to covalent or ionic bonds) have free-moving electrons that transfer thermal energy rapidly through the metal. The *electron fluid* of a conductive metallic solid conducts most of the heat flux through the solid. Phonon flux is still present, but carries less of the energy. Electrons also conduct electric current through conductive solids, and the thermal and electrical conductivities of most metals have about the same ratio. A good electrical conductor, such as copper, also conducts heat well. Thermoelectricity is caused by the interaction of heat flux and electric current. Heat conduction within a solid is directly analogous to diffusion of particles within a fluid, in the situation where there are no fluid currents.

To quantify the ease with which a particular medium conducts, engineers employ the thermal conductivity, also known as the conductivity constant or conduction coefficient, k . In thermal conductivity, k is defined as "the quantity of heat, Q , transmitted in time (t) through a thickness (L), in a direction normal to a surface of area (A), due to a temperature difference (ΔT) [...]". Thermal conductivity is a material *property* that is primarily dependent on the medium's phase, temperature, density, and molecular bonding. Thermal effusivity is a quantity derived from conductivity, which is a measure of its ability to exchange thermal energy with its surroundings.

Steady-state conduction

Steady state conduction is the form of conduction that happens when the temperature difference(s) driving the conduction are constant, so that (after an equilibration time), the spatial distribution of temperatures (temperature field) in the conducting object does not change any further. Thus, all partial derivatives of temperature *with respect to space* may either be zero or have nonzero values, but all derivatives of temperature at any point *with respect to time* are uniformly zero. In steady state conduction, the amount of heat entering any region of an object is equal to amount of heat coming out (if this were not so, the temperature would be rising or falling, as thermal energy was tapped or trapped in a region).

For example, a bar may be cold at one end and hot at the other, but after a state of steady state conduction is reached, the spatial gradient of temperatures along the bar does not change any further, as time proceeds. Instead, the temperature at any given section of the rod remains constant, and this temperature varies linearly in space, along the direction of heat transfer.

In steady state conduction, all the laws of direct current electrical conduction can be applied to "heat currents". In such cases, it is possible to take "thermal resistances" as the analog to electrical resistances. In such cases, temperature plays the role of voltage, and heat transferred per unit time (heat power) is the analog of electric current. Steady state systems can be modelled by networks of such thermal resistances in series and in parallel, in exact analogy to electrical networks of resistors. See purely resistive thermal circuits for an example of such a network.

Transient conduction

In general, during any period in which temperatures change *in time* at any place within an object, the mode of thermal energy flow is termed *transient conduction*. Another term is "non steady-state" conduction, referring to time-dependence of temperature fields in an object. Non-steady-state situations appear after an imposed change in temperature at a boundary of an object. They may also occur with temperature changes inside an object, as a result of a new source or sink of heat suddenly introduced within an object, causing temperatures near the source or sink to change in time.

When a new perturbation of temperature of this type happens, temperatures within the system change in time toward a new equilibrium with the new conditions, provided that these do not change. After equilibrium, heat flow into the system once again equals the heat flow out, and temperatures at each point inside the system no longer change. Once this happens, transient conduction is ended, although steady-state conduction may continue if heat flow continues.

If changes in external temperatures or internal heat generation changes are too rapid for the equilibrium of temperatures in space to take place, then the system never reaches a state of unchanging temperature distribution in time, and the system remains in a transient state.

An example of a new source of heat "turning on" within an object, causing transient conduction, is an engine starting in an automobile. In this case, the transient thermal conduction phase for the entire machine is over, and the steady state phase appears, as soon as the engine reaches steady-state operating temperature. In this state of steady-state equilibrium, temperatures vary greatly from the engine cylinders to other parts of the automobile, but at no point in space within the automobile does temperature increase or decrease. After establishing this state, the transient conduction phase of heat transfer is over.

New external conditions also cause this process: for example the copper bar in the example steady-state conduction experiences transient conduction as soon as one end is subjected to a different temperature from the other. Over time, the field of temperatures inside the bar reach a new steady-state, in which a constant temperature gradient along the bar is finally set up, and this gradient then stays constant in space. Typically, such a new steady state gradient is approached exponentially with time after a new temperature-or-heat source or sink, has been introduced. When a "transient conduction" phase is over, heat flow may still continue at high power, so long as temperatures do not change.

An example of transient conduction that does not end with steady-state conduction, but rather no conduction, occurs when a hot copper ball is dropped into oil at a low temperature. Here, the temperature field within the object begins to change as a function of time, as the heat is removed from the metal, and the interest lies in analyzing this spatial change of temperature within the object over time, until all gradients disappear entirely (the ball has reached the same temperature as the oil). Mathematically, this condition is also approached exponentially; in theory it takes infinite time, but in practice it is over, for all intents and purposes, in a much shorter period. At the end of this process with no heat sink but the internal parts of the ball (which are finite), there is no steady state heat conduction to reach. Such a state never occurs in this situation, but rather the end of the process is when there is no heat conduction at all.

The analysis of non steady-state conduction systems is more complex than that of steady-state systems. If the conducting body has a simple shape, then exact analytical mathematical expressions and solutions may be possible (see heat equation for the analytical approach).^[2] However, most often, because of complicated shapes with varying thermal conductivities within the shape (i.e., most complex objects, mechanisms or machines in engineering) often the application of approximate theories is required, and/or numerical analysis by computer. One popular graphical method involves the use of Heisler Charts.

Occasionally, transient conduction problems may be considerably simplified if regions of the object being heated or cooled can be identified, for which thermal conductivity is very much greater than that for heat paths leading into the region. In this case, the region with high conductivity can often be treated in the lumped capacitance model, as a "lump" of material with a simple thermal capacitance consisting of its aggregate heat capacity. Such regions warm or cool, but show no significant temperature *variation* across their extent, during the process (as compared to the rest of the system). This is due to their far higher conductance. During transient conduction, therefore, the temperature across their conductive regions changes uniformly in space, and as a simple exponential in time. An example of such systems are those that follow Newton's law of cooling during transient cooling (or the reverse during heating). The equivalent thermal circuit consists of a simple capacitor in series with a resistor. In such cases, the remainder of the system with high thermal resistance (comparatively low conductivity) plays the role of the resistor in the circuit.

Relativistic conduction

The theory of relativistic heat conduction is a model that is compatible with the theory of special relativity. For most of the last century, it was recognized that the Fourier equation is in contradiction with the theory of relativity because it admits an infinite speed of propagation of heat signals. For example, according to the Fourier equation, a pulse of heat at the origin would be felt at infinity instantaneously. The speed of information propagation is faster than the speed of light in vacuum, which is physically inadmissible within the framework of relativity.

Quantum conduction

Second sound is a quantum mechanical phenomenon in which heat transfer occurs by wave-like motion, rather than by the more usual mechanism of diffusion. Heat takes the place of pressure in normal sound waves. This leads to a very high thermal conductivity. It is known as "second sound" because the wave motion of heat is similar to the propagation of sound in air.

Fourier's law

The law of heat conduction, also known as Fourier's law, states that the time rate of heat transfer through a material is proportional to the negative gradient in the temperature and to the area, at right angles to that gradient, through which the heat flows. We can state this law in two equivalent forms: the integral form, in which we look at the amount of energy flowing into or out of a body as a whole, and the differential form, in which we look at the flow rates or fluxes of energy locally.

Newton's law of cooling is a discrete analog of Fourier's law, while Ohm's law is the electrical analogue of Fourier's law.

Differential form

The differential form of Fourier's law of thermal conduction shows that the local heat flux density, \vec{q} , is equal to the product of thermal conductivity, k , and the negative local temperature gradient, $-\nabla T$. The heat flux density is the amount of energy that flows through a unit area per unit time.

$$\vec{q} = -k\nabla T$$

where (including the SI units)

\vec{q} is the local heat flux density, $\text{W}\cdot\text{m}^{-2}$
 k is the material's conductivity, $\text{W}\cdot\text{m}^{-1}\cdot\text{K}^{-1}$,
 ∇T is the temperature gradient, $\text{K}\cdot\text{m}^{-1}$.

The thermal conductivity, k , is often treated as a constant, though this is not always true. While the thermal conductivity of a material generally varies with temperature, the variation can be small over a significant range of temperatures for some common materials. In anisotropic materials, the thermal conductivity typically varies with orientation; in this case k is represented by a second-order tensor. In non-uniform materials, k varies with spatial location.

For many simple applications, Fourier's law is used in its one-dimensional form. In the x -direction,

$$q_x = -k \frac{dT}{dx}$$

Integral form

By integrating the differential form over the material's total surface S , we arrive at the integral form of Fourier's law:

$$\frac{\partial Q}{\partial t} = -k \iint_S \nabla T \cdot dS$$

where (including the SI units):

- $\frac{\partial Q}{\partial t}$ is the amount of heat transferred per unit time (in W), and
- dS is an oriented surface area element (in m^2)

The above differential equation, when integrated for a homogeneous material of 1-D geometry between two endpoints at constant temperature, gives the heat flow rate as:

$$\frac{Q}{\Delta t} = -kA \frac{\Delta T}{\Delta x}$$

where

A is the cross-sectional surface area,
 ΔT is the temperature difference between the ends,
 Δx is the distance between the ends.

This law forms the basis for the derivation of the heat equation.

Conductance

Writing

$$U = \frac{k}{\Delta x},$$

where U is the conductance, in $W/(m^2 K)$.

Fourier's law can also be stated as:

$$\frac{\Delta Q}{\Delta t} = UA(-\Delta T).$$

The reciprocal of conductance is resistance, R , given by:

$$R = \frac{1}{U} = \frac{\Delta x}{k} = \frac{A(-\Delta T)}{\frac{\Delta Q}{\Delta t}}.$$

Resistance is additive when several conducting layers lie between the hot and cool regions, because A and Q are the same for all layers. In a multilayer partition, the total conductance is related to the conductance of its layers by:

$$\frac{1}{U} = \frac{1}{U_1} + \frac{1}{U_2} + \frac{1}{U_3} + \dots$$

So, when dealing with a multilayer partition, the following formula is usually used:

$$\frac{\Delta Q}{\Delta t} = \frac{A(-\Delta T)}{\frac{\Delta x_1}{k_1} + \frac{\Delta x_2}{k_2} + \frac{\Delta x_3}{k_3} + \dots}.$$

For heat conduction from one fluid to another through a barrier, it is sometimes important to consider the conductance of the thin film of fluid that remains stationary next to the barrier. This thin film of fluid is difficult to quantify because its characteristics depend upon complex conditions of turbulence and viscosity—but when dealing with thin high-conductance barriers it can sometimes be quite significant.

Intensive-property representation

The previous conductance equations, written in terms of extensive properties, can be reformulated in terms of intensive properties. Ideally, the formulae for conductance should produce a quantity with dimensions independent of distance, like Ohm's Law for electrical resistance, $R = V/I$, and conductance, $G = I/V$.

From the electrical formula: $R = \rho x/A$, where ρ is resistivity, x is length, and A is cross-sectional area, we have $G = kA/x$, where G is conductance, k is conductivity, x is length, and A is cross-sectional area.

For Heat,

$$U = \frac{kA}{\Delta x},$$

where U is the conductance.

Fourier's law can also be stated as:

$$\dot{Q} = U \Delta T,$$

analogous to Ohm's law, $I = V/R$ or $I = VG$.

The reciprocal of conductance is resistance, R , given by:

$$R = \frac{\Delta T}{\dot{Q}},$$

analogous to Ohm's law, $R = V/I$.

The rules for combining resistances and conductances (in series and in parallel) are the same for both heat flow and electric current.

Cylindrical shells

Conduction through cylindrical shells (e.g. pipes) can be calculated from the internal radius, r_1 , the external radius, r_2 , the length, ℓ , and the temperature difference between the inner and outer wall, $T_2 - T_1$.

The surface area of the cylinder is $A_r = 2\pi r\ell$

When Fourier's equation is applied:

$$\dot{Q} = -kA_r \frac{dT}{dr} = -2k\pi r\ell \frac{dT}{dr}$$

and rearranged:

$$\dot{Q} \int_{r_1}^{r_2} \frac{1}{r} dr = -2k\pi\ell \int_{T_1}^{T_2} dT$$

then the rate of heat transfer is:

$$\dot{Q} = 2k\pi\ell \frac{T_1 - T_2}{\ln(r_2/r_1)}$$

the thermal resistance is:

$$R_c = \frac{\Delta T}{\dot{Q}} = \frac{\ln(r_2/r_1)}{2\pi k\ell}$$

and $\dot{Q} = 2\pi k\ell r_m \frac{T_1 - T_2}{r_2 - r_1}$, where $r_m = \frac{r_2 - r_1}{\ln(r_2/r_1)}$. It is important to note that this is the log-mean radius.

Spherical

The conduction through a spherical shell with internal radius, r_1 , and external radius, r_2 , can be calculated in a similar manner as for a cylindrical shell.

The surface area of the sphere is: $A = 4\pi r^2$.

Solving in a similar manner as for a cylindrical shell (see above) produces:

$$\dot{Q} = 4k\pi \frac{T_1 - T_2}{1/r_1 - 1/r_2} = 4k\pi \frac{(T_1 - T_2)r_1 r_2}{r_2 - r_1}$$

Transient thermal conduction

Interface heat transfer

The heat transfer at an interface is considered a transient heat flow. To analyze this problem, the Biot number is important to understand how the system behaves. The Biot number is determined by: $Bi = \frac{hL}{k}$ The heat

transfer coefficient h , is introduced in this formula, and is measured in $\frac{J}{m^2 s K}$. If the system has a Biot number of less than 0.1, the material behaves according to Newtonian cooling, i.e. with negligible temperature gradient within the body. If the Biot number is greater than 0.1, the system behaves as a series solution. The temperature profile in terms of time can be determined by the function can be derived from the equation

$$q = -h \Delta T,$$

which becomes

$$\frac{T - T_f}{T_i - T_f} = \exp\left[\frac{-hAt}{\rho C_p V}\right].$$

The heat transfer coefficient, h , is measured in $\frac{W}{m^2 K}$, and represents the transfer of heat at an interface between two materials. This value is different at every interface, and is an important concept in understanding heat flow at an interface.

The series solution can be analyzed with a nomogram. A nomogram has relative temperature as the y coordinate and the Fourier number, which is calculated by

$$Fo = \frac{\alpha t}{L^2}.$$

The Biot number increases as the Fourier number decreases. There are five steps to determine a temperature profile in terms of time.

1. Calculate the Biot number
2. Determine which relative depth matters, either x or L .
3. Convert time to the Fourier number.
4. Convert T_i to relative temperature with the boundary conditions.
5. Compared required point to trace specified Biot number on the nomogram.

Thermal conduction applications

Splat cooling

Splat cooling is a method for quenching small droplets of molten materials by rapid contact with a cold surface. The particles undergo a characteristic cooling process, with the heat profile at $t = 0$ for initial temperature as the maximum at $x = 0$ and $T = 0$ at $x = -\infty$ and $x = \infty$, and the heat profile at $t = \infty$ for $-\infty \leq x \leq \infty$

as the boundary conditions. Splat cooling rapidly ends in a steady state temperature, and is similar in form to the Gaussian diffusion equation. The temperature profile, with respect to the position and time of this type of cooling, varies with:

$$T(x, t) - T_i = \frac{T_i \Delta X}{2\sqrt{\pi\alpha t}} \exp\left(-\frac{x^2}{4\alpha t}\right)$$

Splat cooling is a fundamental concept that has been adapted for practical use in the form of thermal spraying. The thermal diffusivity coefficient, represented as α , can be written as $\alpha = \frac{k}{\rho C_p}$. This varies according to the material.^{[3][4]}

Metal quenching

Metal quenching is a transient heat transfer process in terms of the Time Temperature Transformation (TTT). It is possible to manipulate the cooling process to adjust the phase of a suitable material. For example, appropriate quenching of steel can convert a desirable proportion of its content of austenite to martensite, creating a very tough product. To achieve this, it is necessary to quench at the "nose" (or eutectic) of the TTT diagram. Since materials differ in their Biot numbers, the time it takes for the material to quench, or the Fourier number, varies in practice.^[5] In steel, the quenching temperature range is generally from 600 °C to 200 °C. To control the quenching time and to select suitable quenching media, it is necessary to determine the Fourier number from the desired quenching time, the relative temperature drop, and the relevant Biot number. Usually, the correct figures are read from a standard nomogram. By calculating the heat transfer coefficient from this Biot number, one can find a liquid medium suitable for the application.^[6]

Zeroth law of thermodynamics

One statement of the so-called zeroth law of thermodynamics is directly focused on the idea of conduction of heat. Baily (1994) writes that "... the zeroth law may be stated:

All diathermal walls are equivalent."^[7]

A diathermal wall is a physical connection between two bodies that allows the passage of heat between them. Baily is referring to diathermal walls that exclusively connect two bodies, especially conductive walls.

This statement of the 'zeroth law' belongs to an idealized theoretical discourse, and actual physical walls may have peculiarities that do not conform to its generality.

For example, the material of the wall must not undergo a phase transition, such as evaporation or fusion, at the temperature at which it must conduct heat. But when only thermal equilibrium is considered and time is not urgent, so that the conductivity of the material does not matter too much, one suitable heat conductor is as good as another. Conversely, another aspect of the zeroth law is that, subject again to suitable restrictions, a given diathermal wall is indifferent to the nature of the heat bath to which it is connected. For example, the glass bulb of a thermometer acts as a diathermal wall whether exposed to a gas or to a liquid, provided they do not corrode or melt it.

These indifferences are amongst the defining characteristics of heat transfer. In a sense, they are symmetries of heat transfer.

See also

- List of thermal conductivities
- Electrical conduction

- Convection diffusion equation
- R-value (insulation)
- Heat pipe
- Fick's law of diffusion
- Relativistic heat conduction
- Churchill–Bernstein equation
- Fourier number
- Biot number
- False diffusion

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External links

- Heat conduction – Thermal-FluidsPedia
- Newton's Law of Cooling by Jeff Bryant based on a program by Stephen Wolfram, Wolfram Demonstrations Project.

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Categories: Concepts in physics | Heat conduction | Heat transfer | Physical quantities

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Water cycle

From Wikipedia, the free encyclopedia

The **water cycle**, also known as the **hydrological cycle** or the **hydrologic cycle**, describes the continuous movement of water on, above and below the surface of the Earth. The mass of water on Earth remains fairly constant over time but the partitioning of the water into the major reservoirs of ice, fresh water, saline water and atmospheric water is variable depending on a wide range of climatic variables. The water moves from one reservoir to another, such as from river to ocean, or from the ocean to the atmosphere, by the physical processes of evaporation, condensation, precipitation, infiltration, surface runoff, and subsurface flow. In doing so, the water goes through different forms: liquid, solid (ice) and vapor.

The water cycle involves the exchange of energy, which leads to temperature changes. For instance, when water evaporates, it takes up energy from its surroundings and cools the environment. When it condenses, it releases energy and warms the environment. These heat exchanges influence climate.

The evaporative phase of the cycle purifies water which then replenishes the land with freshwater. The flow of liquid water and ice transports minerals across the globe. It is also involved in reshaping the geological features of the Earth, through processes including erosion and sedimentation. The water cycle is also essential for the maintenance of most life and ecosystems on the planet.

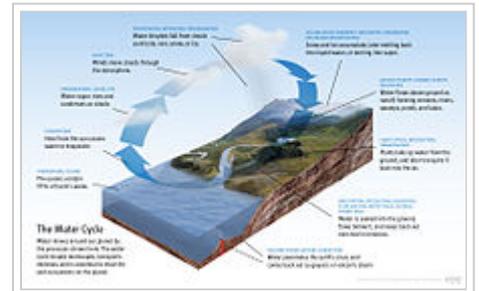
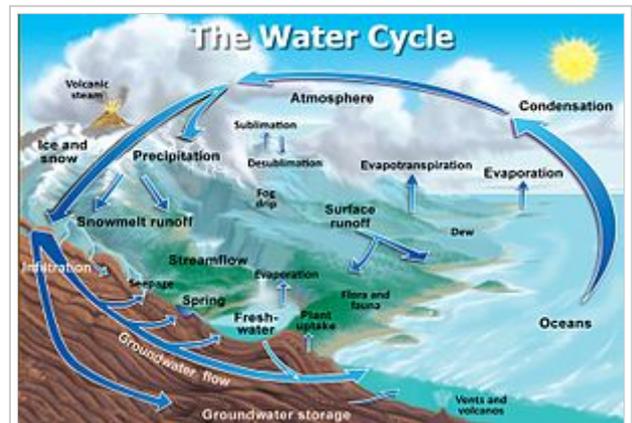


Diagram of the Water Cycle



The water cycle

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Earth's water cycle

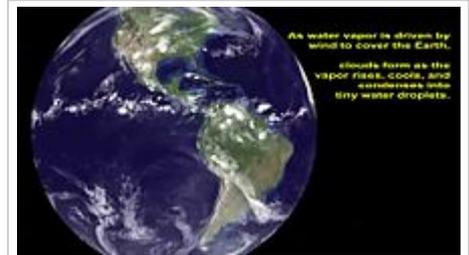


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As the Earth's surface water evaporates, wind moves water in the air from the sea to the land, increasing the amount of freshwater on land.

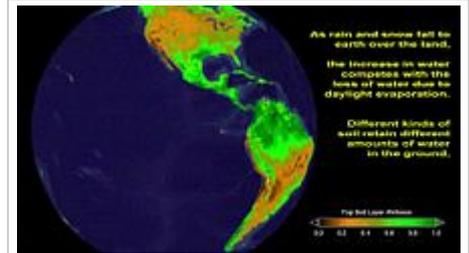
Description

The sun, which drives the water cycle, heats water in oceans and seas. Water evaporates as water vapor into the air. Ice and snow can sublime directly into water vapour. Evapotranspiration is water transpired from plants and evaporated from the soil. The water vapour molecule H_2O has less density compared to the major components of the atmosphere, nitrogen and oxygen, N_2 and O_2 . Due to the significant difference in molecular mass, water vapor in gas form gains height in open air as a result of buoyancy. However, as altitude increases, air pressure decreases and the temperature drops (see Gas laws). The lowered temperature causes water vapour to condense into a tiny liquid water droplet which is heavier than the air, such that it falls unless supported by an updraft. A huge concentration of these droplets over a large space up in the atmosphere become visible as cloud. Fog is formed if the water vapour condenses near ground level, as a result of moist air and cool air collision or an abrupt reduction in air pressure. Air currents move water vapour around the globe, cloud particles collide, grow, and fall out of the upper atmospheric layers as precipitation. Some precipitation falls as snow or hail, sleet, and can accumulate as ice caps and glaciers, which can store frozen water for thousands of years. Most water falls back into the oceans or onto land as rain, where the water flows over the ground as surface runoff. A portion of runoff enters rivers in valleys in the landscape, with streamflow moving water towards the oceans. Runoff and water emerging from the ground (groundwater) may be stored as freshwater in lakes. Not all runoff flows into rivers, much of it soaks into the ground as infiltration. Some water infiltrates deep into the ground and replenishes aquifers, which can store freshwater for long periods of time. Some infiltration stays close to the land surface and can seep back into surface-water bodies (and the ocean) as groundwater discharge. Some groundwater finds openings in the land surface and comes out as freshwater springs. In river valleys and floodplains, there is often continuous water exchange between surface water and ground water in the hyporheic zone. Over time, the water returns to the ocean, to continue the water cycle.



Play media

Water vapor is converted to clouds that bring fresh water to land in the form of rain snow and sleet



Play media

Precipitation falls on the ground, but what happens to that water depends greatly on the geography of the land at any particular place.

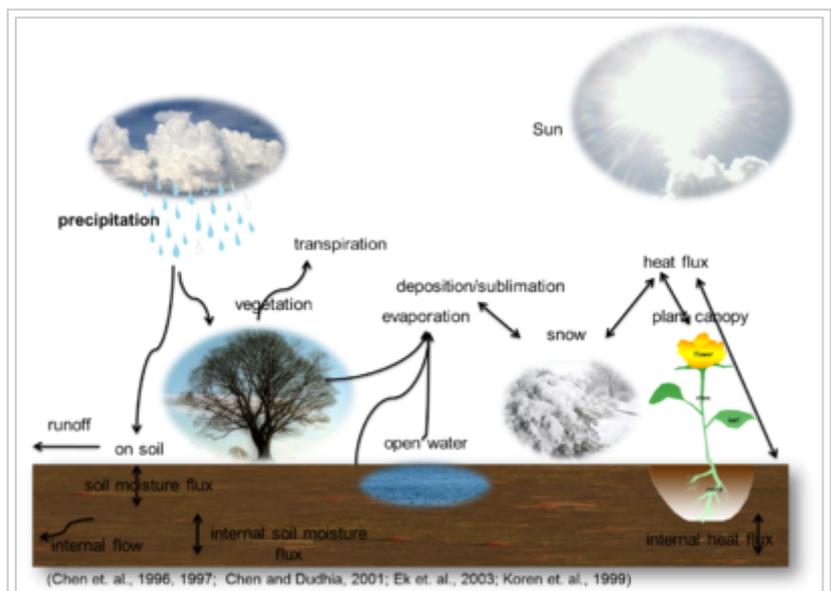
Processes

Precipitation

Condensed water vapor that falls to the Earth's surface. Most precipitation occurs as rain, but also includes snow, hail, fog drip, graupel, and sleet.^[1] Approximately $505,000 \text{ km}^3$ ($121,000 \text{ cu mi}$) of water falls as precipitation each year, $398,000 \text{ km}^3$ ($95,000 \text{ cu mi}$) of it over the oceans.^[2] The rain on land contains $107,000 \text{ km}^3$ ($26,000 \text{ cu mi}$) of water per year and a snowing only $1,000 \text{ km}^3$ (240 cu mi).^[3] 78% of global precipitation occurs over the ocean.^[4]

Canopy interception

The precipitation that is intercepted by plant foliage eventually evaporates back to the atmosphere rather than falling to the ground.



Many different processes lead to movements and phase changes in water

Snowmelt

The runoff produced by melting snow.

Runoff

The variety of ways by which water moves across the land. This includes both surface runoff and channel runoff. As it flows, the water may seep into the ground, evaporate into the air, become stored in lakes or reservoirs, or be extracted for agricultural or other human uses.

Infiltration

The flow of water from the ground surface into the ground. Once infiltrated, the water becomes soil moisture or groundwater.^[5] A recent global study using water stable isotopes, however, shows that not all soil moisture is equally available for groundwater recharge or for plant transpiration.^[6]

Subsurface flow

The flow of water underground, in the vadose zone and aquifers. Subsurface water may return to the surface (e.g. as a spring or by being pumped) or eventually seep into the oceans. Water returns to the land surface at lower elevation than where it infiltrated, under the force of gravity or gravity induced pressures. Groundwater tends to move slowly and is replenished slowly, so it can remain in aquifers for thousands of years.

Evaporation

The transformation of water from liquid to gas phases as it moves from the ground or bodies of water into the overlying atmosphere.^[7] The source of energy for evaporation is primarily solar radiation. Evaporation often implicitly includes transpiration from plants, though together they are specifically referred to as evapotranspiration. Total annual evapotranspiration amounts to approximately 505,000 km³ (121,000 cu mi) of water, 434,000 km³ (104,000 cu mi) of which evaporates from the oceans.^[2] 86% of global evaporation occurs over the ocean.^[4]

Sublimation

The state change directly from solid water (snow or ice) to water vapor.^[8]

Deposition

This refers to changing of water vapor directly to ice.

Advection

The movement of water — in solid, liquid, or vapor states — through the atmosphere. Without advection, water that evaporated over the oceans could not precipitate over land.^[9]

Condensation

The transformation of water vapor to liquid water droplets in the air, creating clouds and fog.^[10]

Transpiration

The release of water vapor from plants and soil into the air. Water vapor is a gas that cannot be seen.

Percolation

Water flows vertically through the soil and rocks under the influence of gravity

Plate tectonics

Water enters the mantle via subduction of oceanic crust. Water returns to the surface via volcanism.

Water cycle thus involves many of the intermediate processes.

Residence times

The *residence time* of a reservoir within the hydrologic cycle is the average time a water molecule will spend in that reservoir (*see adjacent table*). It is a measure of the average age of the water in that reservoir.

Groundwater can spend over 10,000 years beneath Earth's surface before leaving. Particularly old groundwater is called fossil water. Water stored in the soil remains there very briefly, because it is spread thinly across the Earth, and is readily lost by evaporation, transpiration, stream flow, or groundwater recharge. After evaporating, the residence time in the atmosphere is about 9 days before condensing and falling to the Earth as precipitation.

The major ice sheets - Antarctica and Greenland - store ice for very long periods. Ice from Antarctica has been reliably dated to 800,000 years before present, though the average residence time is shorter.^[12]

In hydrology, residence times can be estimated in two ways. The more common method relies on the principle of conservation of mass and assumes the amount of water in a given reservoir is roughly constant. With this method, residence times are estimated by dividing the volume of the reservoir by the rate by which water either enters or exits the reservoir. Conceptually, this is equivalent to timing how long it would take the reservoir to become filled from empty if no water were to leave (or how long it would take the reservoir to empty from full if no water were to enter).

An alternative method to estimate residence times, which is gaining in popularity for dating groundwater, is the use of isotopic techniques. This is done in the subfield of isotope hydrology.

Average reservoir residence times^[11]

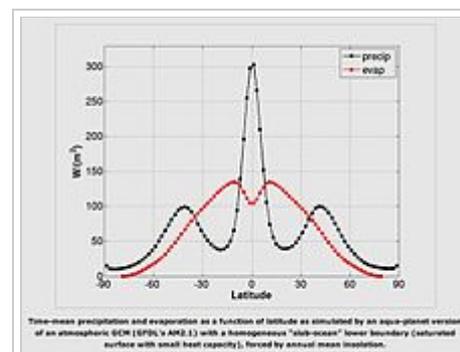
Reservoir	Average residence time
Antarctica	20,000 years
Oceans	3,200 years
Glaciers	20 to 100 years
Seasonal snow cover	2 to 6 months
Soil moisture	1 to 2 months
Groundwater: shallow	100 to 200 years
Groundwater: deep	10,000 years
Lakes (see lake retention time)	50 to 100 years
Rivers	2 to 6 months
Atmosphere	9 days

Changes over time

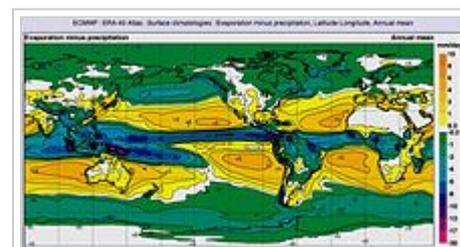
The water cycle describes the processes that drive the movement of water throughout the hydrosphere. However, much more water is "in storage" for long periods of time than is actually moving through the cycle. The storehouses for the vast majority of all water on Earth are the oceans. It is estimated that of the 332,500,000 mi³ (1,386,000,000 km³) of the world's water supply, about 321,000,000 mi³ (1,338,000,000 km³) is stored in oceans, or about 97%. It is also estimated that the oceans supply about 90% of the evaporated water that goes into the water cycle.^[13]

During colder climatic periods more ice caps and glaciers form, and enough of the global water supply accumulates as ice to lessen the amounts in other parts of the water cycle. The reverse is true during warm periods. During the last ice age glaciers covered almost one-third of Earth's land mass, with the result being that the oceans were about 400 ft (122 m) lower than today. During the last global "warm spell," about 125,000 years ago, the seas were about 18 ft (5.5 m) higher than they are now. About three million years ago the oceans could have been up to 165 ft (50 m) higher.^[13]

The scientific consensus expressed in the 2007 Intergovernmental Panel on Climate Change (IPCC) Summary for Policymakers^[14] is for the water cycle to continue to intensify throughout the 21st century, though this does not mean that precipitation will increase in all regions. In subtropical land areas — places that are already relatively dry — precipitation is projected to decrease during the 21st century, increasing the probability of drought. The drying is projected to be strongest near the poleward margins of the subtropics (for example, the Mediterranean Basin, South Africa, southern Australia, and the Southwestern United States). Annual precipitation amounts are expected to increase in near-equatorial regions that tend to be wet in the present climate, and also at high latitudes. These large-scale patterns are present in nearly all of the climate model simulations conducted at several international research centers as part of the 4th Assessment of the IPCC. There is now ample evidence



Time-mean precipitation and evaporation as a function of latitude as simulated by an aqua-planet version of an atmospheric GCM (GFDL's AM2.1) with a homogeneous "slab-ocean" lower boundary (saturated surface with small heat capacity), forced by annual mean insolation.



Global map of annual mean evaporation minus precipitation by latitude-longitude

that increased hydrologic variability and change in climate has and will continue to have a profound impact on the water sector through the hydrologic cycle, water availability, water demand, and water allocation at the global, regional, basin, and local levels.^[15] Research published in 2012 in *Science* based on surface ocean salinity over the period 1950 to 2000 confirm this projection of an intensified global water cycle with salty areas becoming more saline and fresher areas becoming more fresh over the period.^[16]

Fundamental thermodynamics and climate models suggest that dry regions will become drier and wet regions will become wetter in response to warming. Efforts to detect this long-term response in sparse surface observations of rainfall and evaporation remain ambiguous. We show that ocean salinity patterns express an identifiable fingerprint of an intensifying water cycle. Our 50-year observed global surface salinity changes, combined with changes from global climate models, present robust evidence of an intensified global water cycle at a rate of $8 \pm 5\%$ per degree of surface warming. This rate is double the response projected by current-generation climate models and suggests that a substantial (16 to 24%) intensification of the global water cycle will occur in a future 2° to 3° warmer world.^[17]

An instrument carried by the SAC-D satellite launched in June, 2011 measures global sea surface salinity but data collection began only in June, 2011.^{[16][18]}

Glacial retreat is also an example of a changing water cycle, where the supply of water to glaciers from precipitation cannot keep up with the loss of water from melting and sublimation. Glacial retreat since 1850 has been extensive.^[19]

Human activities that alter the water cycle include:

- agriculture
- industry
- alteration of the chemical composition of the atmosphere
- construction of dams
- deforestation and afforestation
- removal of groundwater from wells
- water abstraction from rivers
- urbanization

Effects on climate

The water cycle is powered from solar energy. 86% of the global evaporation occurs from the oceans, reducing their temperature by evaporative cooling.^[20] Without the cooling, the effect of evaporation on the greenhouse effect would lead to a much higher surface temperature of 67°C (153°F), and a warmer planet.

Aquifer drawdown or overdrafting and the pumping of fossil water increases the total amount of water in the hydrosphere, and has been postulated to be a contributor to sea-level rise.^[21]

Effects on biogeochemical cycling

While the water cycle is itself a biogeochemical cycle,^[22] flow of water over and beneath the Earth is a key component of the cycling of other biogeochemicals. Runoff is responsible for almost all of the transport of eroded sediment and phosphorus^[23] from land to waterbodies. The salinity of the oceans is derived from erosion and transport of dissolved salts from the land. Cultural eutrophication of lakes is primarily due to phosphorus, applied in excess to agricultural fields in fertilizers, and then transported overland and down rivers. Both runoff and groundwater flow play significant roles in transporting nitrogen from the land to

waterbodies.^[24] The dead zone at the outlet of the Mississippi River is a consequence of nitrates from fertilizer being carried off agricultural fields and funnelled down the river system to the Gulf of Mexico. Runoff also plays a part in the carbon cycle, again through the transport of eroded rock and soil.^[25]

Slow loss over geologic time

The hydrodynamic wind within the upper portion of a planet's atmosphere allows light chemical elements such as Hydrogen to move up to the exobase, the lower limit of the exosphere, where the gases can then reach escape velocity, entering outer space without impacting other particles of gas. This type of gas loss from a planet into space is known as planetary wind.^[26] Planets with hot lower atmospheres could result in humid upper atmospheres that accelerate the loss of hydrogen.^[27]

History of hydrologic cycle theory

Floating land mass

In ancient times, it was thought that the land mass floated on a body of water, and that most of the water in rivers has its origin under the earth. Examples of this belief can be found in the works of Homer (circa 800 BCE).

Source of rain

In the ancient near east, Hebrew scholars observed that even though the rivers ran into the sea, the sea never became full (Ecclesiastes 1:7). Some scholars conclude that the water cycle was described completely during this time in this passage: "The wind goeth toward the south, and turneth about unto the north; it whirlith about continually, and the wind returneth again according to its circuits. All the rivers run into the sea, yet the sea is not full; unto the place from whence the rivers come, thither they return again" (Ecclesiastes 1:6-7, KJV).^[28] Scholars are not in agreement as to the date of Ecclesiastes, though most scholars point to a date during the time of Solomon, the son of David and Bathsheba, "three thousand years ago,"^[28] there is some agreement that the time period is 962-922 BCE.^[29] Furthermore, it was also observed that when the clouds were full, they emptied rain on the earth (Ecclesiastes 11:3). In addition, during 793-740 BC^[30] a Hebrew prophet, Amos, stated that water comes from the sea and is poured out on the earth (Amos 5:8, 9:6).

Precipitation and percolation

In the Adityahridayam (a devotional hymn to the Sun God) of Ramayana, a Hindu epic dated to the 4th century BC, it is mentioned in the 22nd verse that the Sun heats up water and sends it down as rain. By roughly 500 BCE, Greek scholars were speculating that much of the water in rivers can be attributed to rain. The origin of rain was also known by then. These scholars maintained the belief, however, that water rising up through the earth contributed a great deal to rivers. Examples of this thinking included Anaximander (570 BCE) (who also speculated about the evolution of land animals from fish^[31]) and Xenophanes of Colophon (530 BCE).^[32] Chinese scholars such as Chi Ni Tzu (320 BC) and Lu Shih Ch'un Ch'iu (239 BCE) had similar thoughts.^[33] The idea that the water cycle is a closed cycle can be found in the works of Anaxagoras of Clazomenae (460 BCE) and Diogenes of Apollonia (460 BCE). Both Plato (390 BCE) and Aristotle (350 BCE) speculated about percolation as part of the water cycle.

Precipitation alone

In the Biblical Book of Job, dated between 7th and 2nd centuries BCE,^[29] there is a description of precipitation in the hydrologic cycle,^[28] "For he maketh small the drops of water: they pour down rain according to the vapour thereof; Which the clouds do drop and distil upon man abundantly" (Job 36:27-28, KJV). Also found in the book of Ecclesiastes "All the rivers flow into the sea, Yet the sea is not full. To the place where the rivers flow, There they flow again." (Ecclesiastes 1:7)

Up to the time of the Renaissance, it was thought that precipitation alone was insufficient to feed rivers, for a complete water cycle, and that underground water pushing upwards from the oceans were the main contributors to river water. Bartholomew of England held this view (1240 CE), as did Leonardo da Vinci (1500 CE) and Athanasius Kircher (1644 CE).

The first published thinker to assert that rainfall alone was sufficient for the maintenance of rivers was Bernard Palissy (1580 CE), who is often credited as the "discoverer" of the modern theory of the water cycle. Palissy's theories were not tested scientifically until 1674, in a study commonly attributed to Pierre Perrault. Even then, these beliefs were not accepted in mainstream science until the early nineteenth century.^[34]

See also

- Bioprecipitation
- Drought
- Ecohydrology
- Flood
- Moisture advection
- Moisture recycling
- Planetary boundaries
- Water use

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External links

- The Water Cycle, United States Geological Survey
- The Water Cycle for Kids, United States Geological Survey
- The water cycle, from *Dr. Art's Guide to the Planet*.
- Water cycle slideshow, 1 Mb Flash multilingual animation highlighting the often-overlooked evaporation from bare soil, from managingwholes.com.
- Will the wet get wetter and the dry drier? - Climate research summary from NOAA Geophysical Fluid Dynamics Laboratory including text, graphics, and animations



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Wien's displacement law

From Wikipedia, the free encyclopedia

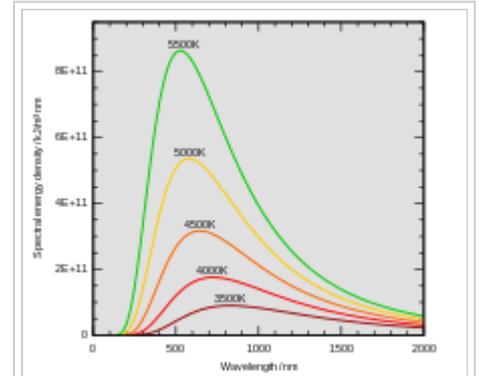
Wien's displacement law states that the black body radiation curve for different temperatures peaks at a wavelength inversely proportional to the temperature. The shift of that peak is a direct consequence of the Planck radiation law which describes the spectral brightness of black body radiation as a function of wavelength at any given temperature. However it had been discovered by Wilhelm Wien several years before Max Planck developed that more general equation, and describes the entire shift of the spectrum of black body radiation toward shorter wavelengths as temperature increases.

Formally, Wien's displacement law states that the spectral radiance of black body radiation per unit wavelength, peaks at the wavelength λ_{\max} given by:

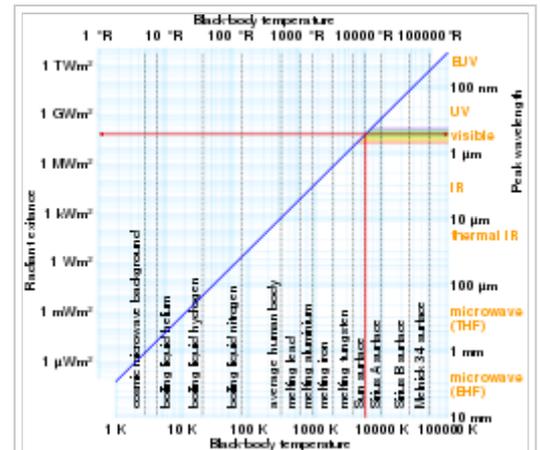
$$\lambda_{\max} = \frac{b}{T}$$

where T is the absolute temperature in kelvins. b is a constant of proportionality called *Wien's displacement constant*, equal to $2.897\,7729(17) \times 10^{-3} \text{ m}\cdot\text{K}^{[1]}$, or more conveniently to obtain wavelength in micrometers, $b \approx 2900 \text{ }\mu\text{m}\cdot\text{K}$. If one is considering the peak of black body emission per unit frequency or per proportional bandwidth, one must use a different proportionality constant. However the form of the law remains the same: the peak wavelength is inversely proportional to temperature (or the peak frequency is directly proportional to temperature).

Wien's displacement law may be referred to as "Wien's law", a term which is also used for the Wien approximation.



Black body radiation as a function of wavelength for various absolute temperatures. Each curve is seen to peak at a somewhat different wavelength; Wien's law describes the shift of that peak in terms of temperature.



Log-log graphs of peak emission wavelength and radiant exitance vs black-body temperature – red arrows show that 5780 K black bodies have 501 nm peak wavelength and 63.3 MW/m² radiant exitance

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Examples

Wien's displacement law is relevant to some everyday experiences:

- A piece of metal heated by a torch first becomes "red hot" as the very longest visible wavelengths appear red, then becomes more orange-red as the temperature is increased, and at very high temperatures would

be described as "white hot" as shorter and shorter wavelengths come to predominate the black body emission spectrum. Before it had even reached the red hot temperature, the thermal emission was mainly at longer infrared wavelengths which are not visible; nevertheless that radiation could be felt as it warms one's nearby skin.

- One easily observes changes in the color of an incandescent light bulb (which produces light through thermal radiation) as the temperature of its filament is varied by a light dimmer. As the light is dimmed and the filament temperature decreases, the distribution of color shifts toward longer wavelengths and the light appears redder, as well as dimmer.
- It is easy to calculate that a wood fire at 1500 K puts out peak radiation at about 2000 nm. 98% of its radiation is beyond 1000 nm and only a tiny proportion at visible wavelengths. Consequently, a campfire can keep one warm but is a poor source of visible light.
- The effective temperature of the Sun is 5778 K. Using Wien's law, one finds a peak emission per nanometer (of wavelength) at a wavelength of about 500 nm in the green portion of the spectrum near the peak sensitivity of the human eye.^{[2][3]} On the other hand, in terms of power per unit optical frequency, the Sun's peak emission is at 343 THz or a wavelength of 883 nm in the near infrared. In terms of power per percentage bandwidth, the peak is at about 635 nm, a red wavelength. Regardless of how one wants to plot the spectrum, about half of the sun's radiation is at wavelengths shorter than 710 nm, about the limit of the human vision. Of that, about 12% is at wavelengths shorter than 400 nm, ultraviolet wavelengths which cannot be seen. It can be appreciated that a rather large amount of the Sun's radiation falls in the fairly small visible spectrum.
- The preponderance of emission in the visible range, however, is not the case in most stars. The hot supergiant Rigel emits 60% of its light in the ultraviolet, while the cool supergiant Betelgeuse emits 85% of its light at infrared wavelengths. With both stars prominent in the constellation of Orion, one can easily appreciate the color difference between the blue-white Rigel ($T = 12100$ K) and the red Betelgeuse ($T \approx 3300$ K). While few stars are as hot as Rigel, stars cooler than the sun or even as cool as Betelgeuse are very commonplace.
- Mammals with a skin temperature of about 300 K emit peak radiation at around $10 \mu\text{m}$ in the far infrared. This is therefore the range of infrared wavelengths that pit viper snakes and passive IR cameras must sense.
- When comparing the apparent color of lighting sources (including fluorescent lights, LED lighting, computer monitors, and photoflash), it is customary to cite the color temperature. Although the spectra of such lights are not accurately described by the black body radiation curve, a color temperature is quoted for which black body radiation would most closely match the subjective color of that source. For instance, the blue-white fluorescent light sometimes used in an office may have a color temperature of 6500 K, whereas the reddish tint of a dimmed incandescent light may have a color temperature (and an actual filament temperature) of 2000 K. Note that the informal description of the former (bluish) color as "cool" and the latter (reddish) as "warm" is exactly opposite the actual temperature change involved in black body radiation.



The color of a star is determined by its temperature, according to Wien's law. In the constellation of Orion, one can compare Betelgeuse ($T \approx 3300$ K, upper left), Rigel ($T = 12100$ K, bottom right), Bellatrix ($T = 22000$ K, upper right), and Mintaka ($T = 31800$ K, rightmost of the 3 "belt stars" in the middle).

Discovery

The law is named for Wilhelm Wien, who derived it in 1893 based on a thermodynamic argument.^[4] Wien considered adiabatic expansion of a cavity containing waves of light in thermal equilibrium. He showed that under slow expansion or contraction, the energy of light reflecting off the walls changes in exactly the same way as the frequency. A general principle of thermodynamics is that a thermal equilibrium state, when

expanded very slowly stays in thermal equilibrium. The adiabatic principle allowed Wien to conclude that for each mode, the adiabatic invariant energy/frequency is only a function of the other adiabatic invariant, the frequency/temperature. A modern variant of Wien's derivation can be found in the textbook by Wannier.^[5]

The consequence is that the shape of the black body radiation function (which was not yet understood) would shift proportionally in frequency (or inversely proportionally in wavelength) with temperature. When Max Planck later formulated the correct black body radiation function it did not include Wien's constant b explicitly. Rather, Planck's constant h was created and introduced into his new formula. From Planck's constant h and the Boltzmann constant k , Wien's constant b can be obtained.

Frequency-dependent formulation

For spectral flux considered per unit frequency $d\nu$ (in hertz), Wien's displacement law describes a peak emission at the optical frequency ν_{\max} given by:

$$\nu_{\max} = \frac{\alpha}{h} kT \approx (5.879 \times 10^{10} \text{ Hz/K}) \cdot T$$

where $\alpha \approx 2.821439\dots$ is a constant resulting from the numerical solution of the maximization equation, k is the Boltzmann constant, h is the Planck constant, and T is the temperature (in kelvins). This frequency does not correspond to the wavelength from the earlier formula which considered the peak emission per unit wavelength.

Derivation from Planck's Law

Planck's law for the spectrum of black body radiation predicts the Wien displacement law and may be used to numerically evaluate the constant relating temperature and peak wavelength (or frequency). According to one form of that law, the black body spectral radiance (power per emitting area per solid angle *per unit wavelength*) is given by:

$$u_{\lambda}(\lambda, T) = \frac{2hc^2}{\lambda^5} \frac{1}{e^{hc/\lambda kT} - 1}.$$

Differentiating $u(\lambda, T)$ with respect to λ and setting the derivative equal to zero gives

$$\frac{\partial u}{\partial \lambda} = 2hc^2 \left(\frac{hc}{kT\lambda^7} \frac{e^{hc/\lambda kT}}{(e^{hc/\lambda kT} - 1)^2} - \frac{1}{\lambda^6} \frac{5}{e^{hc/\lambda kT} - 1} \right) = 0,$$

which can be simplified to give

$$\frac{hc}{\lambda kT} \frac{e^{hc/\lambda kT}}{e^{hc/\lambda kT} - 1} - 5 = 0.$$

By defining:

$$x \equiv \frac{hc}{\lambda kT},$$

the equation becomes one in that single variable:

$$\frac{x e^x}{e^x - 1} - 5 = 0.$$

The numerical solution to this equation is^[note 1] $x = 4.965114231$.

Solving for the wavelength λ in units of nanometers, and using kelvin for the temperature yields:

$$\lambda_{\max} = \frac{hc}{x} \frac{1}{kT} = \frac{2.89776829 \times 10^6 \text{ nm} \cdot \text{K}}{T}.$$

The form of Wien's displacement law in terms of maximum radiance per unit *frequency* is derived using similar methods, but starting with the form of Planck's law expressed in those terms rather than wavelength. The effective result is to substitute 3 for 5 in the equation for the peak wavelength. This is solved with $x = 2.821439372$.

Using the value 4 in this equation solves for the wavelength of the peak in the spectral radiance expressed in radiance *per proportional bandwidth*, perhaps a fairer way of presenting "wavelength of peak emission." That is solved as $x = 3.920690395$. The important point of Wiens law, however, is that *any* such wavelength marker, including the median wavelength (or the wavelength below which a specified percentage of the emission occurs) is proportional to the reciprocal of temperature.

See also

- Wien approximation
- Emissivity
- Sakuma–Hattori equation
- Stefan–Boltzmann law
- Thermometer

Notes

1. The equation $xe^x / (e^x - 1) = n$ cannot be solved in terms of elementary functions. It can be solved in terms of Lambert's product log function as $x = n + W(-ne^{-n})$, but the exact solution is not important in this derivation.

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External links

- Eric Weisstein's World of Physics

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Zenith

From Wikipedia, the free encyclopedia

The **zenith** is an imaginary point directly "above" a particular location, on the imaginary celestial sphere. "Above" means in the vertical direction opposite to the apparent gravitational force at that location. The opposite direction, i.e. the direction in which gravity pulls, is toward the nadir. The zenith is the "highest" point on the celestial sphere (meaning it is the farthest up from the gravitational force).

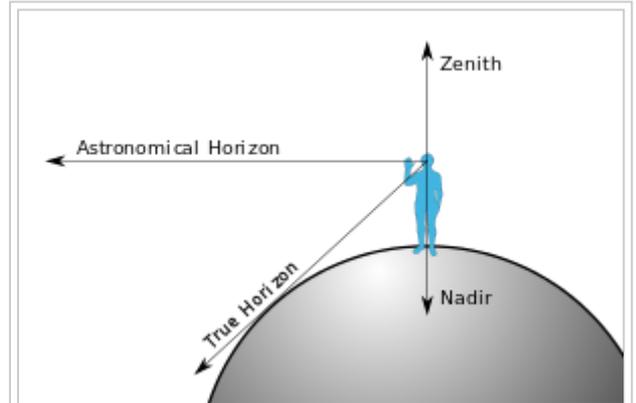


Diagram showing the relationship between the zenith, the nadir, and different types of horizon. Note that the zenith is opposite the nadir

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- 2 Relevance and use
- 3 See also
- 4 References



Look up **zenith** in Wiktionary, the free dictionary.

Origin

The word "zenith" derives from an inaccurate reading of the Arabic expression **سمت الرأس** (*samt ar-ra's*) proposed by ancient Persian astronomers, meaning "direction of the head" or "path above the head", by Medieval Latin scribes in the Middle Ages (during the 14th century), possibly through Old Spanish.^[1] It was reduced to 'samt' ("direction") and miswritten as 'senit'/'cenit', as the "m" was misread as an "ni". Through the Old French 'cenith', 'zenith' first appeared in the 17th century.^{[2][3]}

Relevance and use

The term *zenith* is sometimes used to refer to the highest point, way or level reached by a celestial body during its apparent orbit around a given point of observation.^[4] This sense of the word is often used to describe the location of the Sun ("The sun reached its zenith..."), but to an astronomer the sun does not have its own zenith, and is at the zenith only if it is directly overhead.

In a scientific context, the zenith is the direction of reference for measuring the **zenith angle**, the angle between a direction of interest (e.g., a star) and the local zenith.

In astronomy, the altitude in the horizontal coordinate system and the zenith angle are complementary angles, with the horizon perpendicular to the zenith. The astronomical meridian is also determined by the zenith, and is defined as a circle on the celestial sphere that passes through the zenith, nadir, and the celestial poles. A zenith telescope is a type of telescope designed to point straight up at or near the zenith, and used for precision measurement of star positions, to simplify telescope construction, or both. The NASA Orbital Debris Observatory and the Large Zenith Telescope are both zenith telescopes since the use of liquid mirrors meant these telescopes could only point straight up.



Shadows of trees when the sun is directly overhead (at the zenith). This happens at the solar noon if the tree's latitude equals the sun's declination at that moment.

On the International Space Station, *zenith* and *nadir* are used instead of *up* and *down*, referring to directions within and around the station, relative to the earth.

See also

- Azimuth
- Geodesy
- History of geodesy
- Keyhole problem
- Midheaven
- Subsolar point
- Vertical deflection
- Horizontal coordinate system

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