

Environmental Geology

Malaspina University-College GEOL-312

Unit 1 Review, clay minerals, atmosphere and oceans and climate change

1.1 Review of physical geology

Some of the fundamentals of physical geology will be reviewed in the following few pages, beginning with plate tectonics and ending with minerals and rocks. If you feel that you need additional information on any of the topics covered here you should find an introductory geology textbook and read some of the relevant chapters.

Plate tectonics

Plate tectonics is the theory that the earth's crust is made up of a number of semi-rigid plates, and that these plates are continually moving. It took many decades for geologists to accept plate tectonics, but with an increasing understanding of the processes involved, we are constantly learning more about how it applies to virtually every aspect of geology.

Significance of
Plate Tectonic
theory

To understand plate tectonics it is necessary first to understand the internal structure of the earth [Keller, Figure 2.2¹]. The main components of the earth's interior are as follows:

Earth's internal
structure

The **core** is largely made up of iron and nickel. The inner core is solid, while the outer core is liquid. Convective motion within the outer core is responsible for the earth's magnetic field.

The **mantle** is composed primarily of ferromagnesian silicate minerals. While this material is essentially solid, it is slowly convecting in response to intense forces. The zone from a depth of 100 km to around 250 km - known as the asthenosphere - is more liquid than the rest. The uppermost part of the mantle is essentially solid, and forms part of the lithosphere (along with the crust).

Primarily as a result of the radioactive decay of potassium, thorium and uranium, there is a tremendous amount of heat generated within the earth. This heat is the driving force behind convection currents in the core and mantle- and convection is an important part of the plate motion mechanism.

Internal heat

The **crust** has two distinctive characteristics. In continental areas the crust is felsic (or sialic) in composition, and is between 30 and 40 km thick. In oceanic areas the crust is mafic (simatic) in composition, and is around 5 km thick. The earth's plates are

Oceanic and
continental crust

¹ In these notes all references to **Environmental Geology**, 8th Ed. (Keller) are enclosed in [square brackets].

comprised of crust and rigid upper mantle material - the lithosphere - and they rest on the asthenosphere, where the mantle is partially melted and very susceptible to deformation. There are seven major plates, and numerous minor plates [Figure 2.3].

There are three types of boundaries between plates, namely: divergent, convergent, and transform [Figure 2.4]. Plate boundaries are further classified on the basis of the type of crust, continental or oceanic, which is present on either side of the boundary.

Plate boundaries

Divergent boundaries are spreading boundaries, where new oceanic crust is created from molten mantle material. Most are situated at the oceanic-ridges. Spreading is partly caused by the convective movement within the mantle, which has the effect of pulling the plates apart [Figure 2.1]. Magma (molten rock) from the mantle pushes up to fill the cracks left by spreading. A variety of volcanic rocks are created at the spreading centre (pillow lavas where magma is pushed out into sea-water, dykes beneath that, and gabbro at the base of the oceanic crust).

Divergent boundaries

Convergent boundaries are of three types depending on what type of crust is present on either side of the boundary (ie. ocean-ocean, ocean-continent, continent-continent).

Convergent boundaries

a) **Ocean-Ocean** At an ocean-ocean convergent boundary one of the plates (ocean crust and lithospheric mantle) will be pushed under, or subducted under the other. The subducted lithosphere will descend into the hot mantle, and will start to melt, particularly along the upper surface that is covered with a layer of sediments. Examples of ocean-ocean convergent zones are: subduction of the Pacific plate south of Alaska (Aleutian Islands), subduction of the India-Australian plate south of Indonesia; and subduction of the Atlantic Plate beneath the Caribbean Plate.

b) **Ocean-continent** At an ocean-continent convergent boundary the oceanic plate will be pushed under the continental plate in the same manner as an ocean-ocean collision. Similar geological features apply [Figures 2.4 and 2.10]. Examples are: Nazca plate under South America (which has created the Andes Range), Juan de Fuca plate under North America (creating mountains like Garibaldi, Baker, St. Helens, Ranier, Hood and Shasta).

c) **Continent-continent** A continent-continent collision occurs when a continent or large island has been moved along with oceanic crust (which was being subducted under another continent), and then collides with that other continent. The colliding continental material will not be subducted because it is too light, but the convection continues, so the root of the oceanic plate breaks off and is absorbed into the mantle. There is tremendous deformation of the pre-existing continental rocks, and creation of mountains from that rock. Examples are: the Indo-Australian plate into the Eurasian plate, to create the Himalaya Mountains, and the African plate into the Eurasian plate, to create the Alps in Europe and the Zagros Mts. in Iran).

Transform boundaries exist where one plate slides past another, without production or destruction of crustal material. Most transform faults connect segments of mid-ocean ridges and are thus ocean-ocean boundaries. Some transform faults connect continental parts of plates. Examples are the San Andreas fault, which connects the Juan de Fuca ridge with the Gulf of California ridge and the Queen Charlotte Fault [Figure 2.6]. Transform faults do not just connect divergent boundaries. For instance

Transform boundaries

the convergent boundary beneath the Himalayas is connected to the subduction zone beneath Indonesia by a transform fault. On the Anatolian fault, along the northern edge of Turkey, the Anatolian Plate is being pushed westward with respect to the Eurasian Plate.

Minerals

Mineral properties

A mineral is a naturally occurring, inorganic solid with a definite chemical composition and a regular internal crystal structure. In essence, minerals are molecules, made up of atoms. Rocks are made up of minerals.

All minerals are characterized by specific three-dimensional patterns that are known as lattices or crystal structures. These structures range from the simple cubic pattern of halite (NaCl), to the very complex patterns of some silicate minerals. Two minerals, such as graphite and diamond, may have the same composition, but very different crystal structures and properties.

The mode of formation for most of the minerals that make up the rocks around us is through the cooling of a magma - a body of molten rock. Some geological materials are liquids at the high temperatures that exist deep within the crust and in the upper part of the mantle. As these magmas rise up through the crust they cool enough so that the minerals within them start to crystallize. A second important mechanism for mineral formation is through precipitation from aqueous solution. Water, with dissolved solids, flows through cracks in a body of rock. When the conditions (such as temperature, pressure, chemistry) change some minerals may no longer remain dissolved in the water and will precipitate. Minerals also precipitate from bodies of water. Calcite, for example, can precipitate directly from sea-water, or with the help of shell-building organisms. A third important mechanism for mineral formation is through metamorphism. In response to increasing heat, for example, garnet will form within a shaly rock that had no garnet to begin with.

Mineral formation

Minerals are classified on the basis of their predominant anion or anion group. Some examples include oxides, sulphides, carbonates and silicates. Silicates are by far the predominant group, and will be discussed later. Some examples of minerals from the different mineral groups are given in the table below.

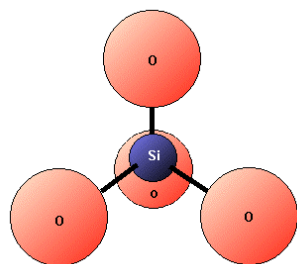
Mineral classification

Group	Mineral name	Composition
OXIDES	Magnetite	Fe ₃ O ₄
	Hematite	Fe ₂ O ₃
SULPHIDES	Pyrite	FeS ₂
	Chalcopyrite	CuFeS ₂
	Galena	PbS
CARBONATES	Calcite	CaCO ₃
	Dolomite	(Ca,Mg)CO ₃
SULPHATES	Gypsum	CaSO ₄ .H ₂ O
NATIVE	Graphite	C

Some important non-silicate minerals and their compositions

The vast majority of the minerals that make up the earth's crust are silicate minerals -

such as quartz, feldspar, mica, amphibole, pyroxene, olivine, and a variety of clay minerals (see table below). The building block of all of these minerals is the silica tetrahedron, a combination of four oxygen atoms and one silicon atom.



In silicate minerals these tetrahedra are arranged and linked together in a variety of ways. The simplest silicate structure, that of olivine, is composed of single isolated tetrahedra bonded to iron and magnesium ions. In pyroxene the silica tetrahedra are linked together in single chain, while in amphibole minerals (such as hornblende) the silica tetrahedra are linked in double chains. In micas the silica tetrahedra are arranged in continuous sheets. Bonding within the sheets is strong but that between sheets is relatively weak, and this accounts for the well developed one-directional cleavage of micaceous minerals.

Silicate minerals

Apart from muscovite, biotite and chlorite, there are numerous other "sheet" silicates that usually exist as clay-sized fragments (ie. less than 0.004 mm). These include kaolinite, illite and smectite, and several other minerals. The clay minerals are of particular importance to various aspects of environmental geology, and they will be discussed in more detail later.

Sheet silicates

Silica tetrahedra are bonded in three-dimensional frameworks in both the feldspars and quartz. The three feldspar "end-members" are orthoclase (potassium only), and two types of plagioclase: albite (sodium only) and anorthite (calcium only). There is a continuous range of compositions (called "solid-solution") between albite and anorthite because the calcium and sodium ions are almost identical in size.

Feldspars

In quartz the bonding between silica tetrahedra is such that the ratio of oxygen to silicon is 2:1 and the charge is balanced. There are no additional cations. The hardness and lack of cleavage in quartz are related to the fact that all of the bonds are - in part - strong covalent bonds.

Mineral name	Silicate type	Composition
Olivine	isolated	(Mg,Fe) silicate
Pyroxene	single chain	(Mg,Fe,Ca) silicate
Amphibole	double chain	(Mg,Fe,Na,K,Ca)(OH) silicate
Biotite	sheet	(K,Mg,Fe,)(Al)(OH) silicate
Garnet	isolated	(Mg,Fe,Ca,Mn,Al) silicate
Muscovite	sheet	(K)(Al)(OH) silicate
Plagioclase feldspar	framework	CaAl ₂ Si ₂ O ₈ to NaAlSi ₃ O ₈
K-feldspar	framework	KAlSi ₃ O ₈
Quartz	framework	SiO ₂
Kaolin	sheet (clay)	Al, OH silicate
Illite	sheet (clay)	K,Al,OH silicate
Smectites (eg. montmorillonite)	sheet (clay)	(Ca,Na,Al,Mg,Fe)(OH)silicate(H ₂ O)

Common silicate minerals and their compositions

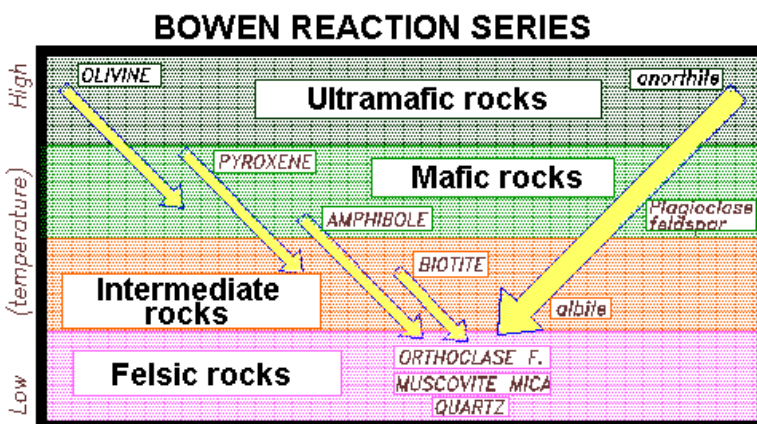
Magmas and Rocks

Rocks

Magmas can have quite widely varying compositions, but they are all made up largely of only eight elements, in order of importance: oxygen, silicon, aluminum, iron, calcium, sodium, magnesium and potassium. Magmas derived from recycled crustal material are dominated by oxygen, silicon and aluminum, sodium and potassium. Magmas derived from the mantle material beneath the crust have higher levels of iron, magnesium and calcium, but they are still likely to be dominated by oxygen and silicon. All magmas also have smaller amounts of elements like hydrogen, nitrogen, carbon and sulphur, which are emitted as water-vapour, carbon dioxide, nitrogen gas, hydrogen sulphide and other gases during an eruption.

At temperatures of several thousands of degrees magma will be entirely liquid, but as the temperature drops, usually because the magma is moving higher into the crust, crystals will start to form. The sequence in which minerals crystallize from a magma - in response to cooling - is known as the Bowen Reaction Series. Once a mineral has crystallized it may continue to react with the remaining magma to form different minerals. For example, olivine crystals will combine with SiO₂ left in the magma to form pyroxene.

Cooling of magmas



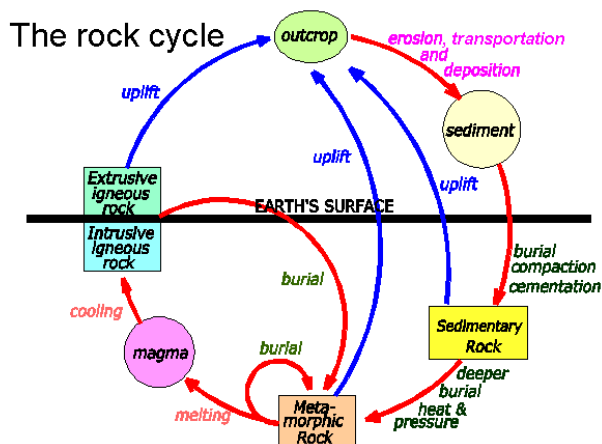
Order of mineral crystallization and reaction according to Bowen's reaction series

The broad compositional classes for igneous rocks are felsic - for rocks that are dominated by non-ferromagnesian minerals (silicate minerals without iron and magnesium) and mafic - for rocks that are dominated by ferromagnesian minerals. Felsic rocks are also known as SIALIC - which refers to the predominance of silica and alumina, while mafic rocks are known as SIMATIC - referring to the predominance of magnesium and iron.

Felsic and mafic igneous rocks

Felsic magmas are more viscous than mafic magmas because they have a higher silica content, and it is the polymerization of silica tetrahedra which promotes high viscosity. Felsic magmas also have higher gas contents than mafic magmas.

The inter-relationships between rock types can be summarized on the rock cycle diagram.



The rock cycle - relationships amongst the three major rock types

Some of the important processes that take place in the transformation of one rock type to another are as follows:

- Uplift and/or exposure - A body of rock cannot be weathered until it is exposed at surface, commonly through a combination of uplift (by tectonic forces) and exposure (the removal of overlying rocks).
- Weathering - Solid rock outcrops are broken down into pieces by a combination of chemical and mechanical processes. The products are small rock and mineral fragments. The most likely minerals to survive this weathering are quartz and feldspar - especially quartz. Other minerals will be either completely dissolved, or altered into clay minerals.
- Transportation and deposition - The products of weathering will be transported by water, wind, ice and gravity. Water is by far the most important, and the velocity of that water will control what size of material can be transported. Where the water slows down some material will be deposited. Once the material reaches a lake or the ocean, all will eventually be deposited, but the finer material will travel further out into the still water.
- Burial and lithification - When sediments are buried by more sediments they begin to be compressed. They lose much of their water content, and eventually become lithified as the individual grains are cemented together.
- Metamorphism - With continual burial the temperature and pressure increase and the existing minerals (especially the clays) become unstable in favour of new minerals - such as mica or garnets. The new minerals grow with their long axes parallel to the least amount of pressure - and hence the rocks have foliation. At extreme burial depths some of the material will begin to melt.

1.2 Clay minerals

Clay minerals

Clay is broadly defined as any mineral material consistently less than 0.004 mm in diameter. This could include finely ground up quartz, feldspar, calcite, hematite or any other mineral, however most clay deposits are dominated by the *clay minerals*,

which belong to the sheet silicates or phyllosilicates (from the Greek word *phyllo* meaning leaf).

A **clay mineral** - in the strict sense - is a finely crystalline sheet silicate with hydroxyl ions, and in some cases with water as part of the structure. By sheet silicate we mean that the silica tetrahedra are arranged in flat sheets - with strong covalent bonding within the sheets, and that these sheets are arranged in layers - where the bonding between the layers is relatively weak. A hydroxyl ion is an oxygen-hydrogen pair (OH⁻), and these form a major part of all clay minerals. Some clay minerals also have H₂O as part of the structure, or in some cases H₂O is simply attached onto the structure.

Most of the clay present in the upper part of the crust has formed as a result of weathering of other silicate minerals, primarily feldspars, micas, pyroxene and amphibole. The reactions involved are hydration reactions, something like the following reaction of **potassium feldspar** plus water and carbon dioxide to **kaolin**² (from Hounslow, 1995):

Formation of clay minerals



In simple terms, K-feldspar reacts with water and carbon dioxide to form kaolin plus carbonic acid. The potassium and some of the silicon that were originally present in the feldspar, are removed in solution. Clay minerals can also be formed when hot waters (known as *hydrothermal solutions*) circulate through a body of rock. As is the case for weathering, the hot solutions lead to alteration of pre-existing minerals.

Hydrothermal alteration

Hydrothermal solutions are often also associated with the deposition of metal deposits (such as the porphyry copper deposits of British Columbia) and the surrounding clay-mineral halos can be an important guide to the possible existence of such deposits.

The clay minerals, which are formed during weathering of other silicate minerals, are soft and easily eroded into tiny fragments and then transported. They accumulate as sediments in low-energy deposition environments (eg. off-shore), sediments that are eventually turned into shale and clay-rich sandstone.

Erosion of clay minerals

It is useful to understand some of the properties of clay minerals, as they have many important implications for environmental geology. They play a role in a great variety of processes, from the origin of life to the causes and effects of earthquakes, to the permeability of rocks to groundwater.

Clay mineral properties

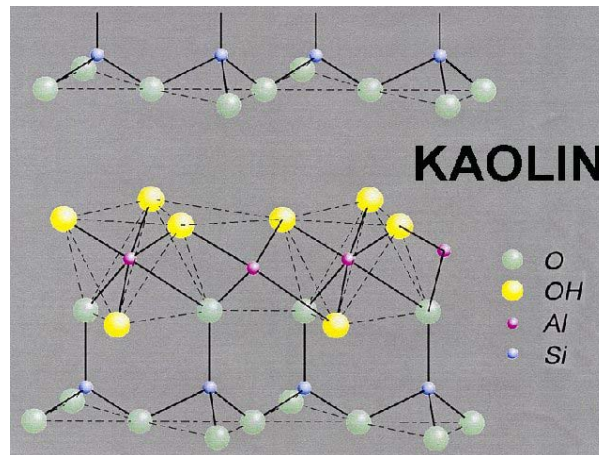
The first important feature of clay minerals is based on the characteristics of the bonding, because the sheets themselves are strongly bonded internally, but only weakly bonded one to another. The sheets that make up a clay mineral grain have a tendency to slide with respect to each other, and the result is that clay mineral masses

Clay mineral bonding

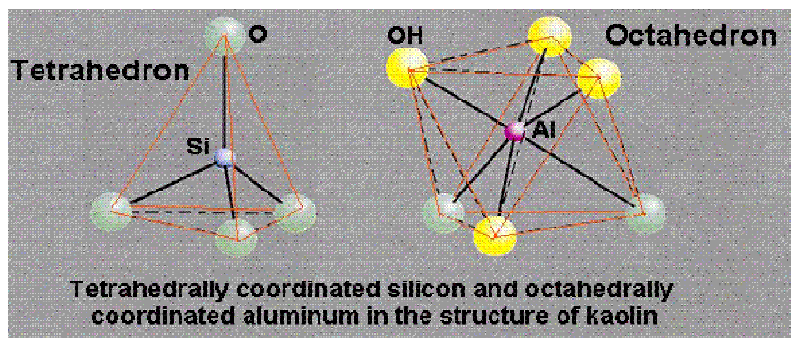
² There are several different forms of the mineral kaolin. The most well known is kaolinite, which is the typical product of weathering of feldspars. Others include halloysite, dickite and nacrite. The latter two are higher-temperature forms.

tend to be soft and plastic, and not very strong.

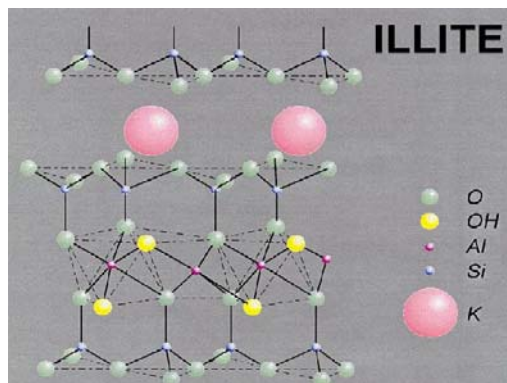
The simplest clay mineral is kaolin. Each “sheet” within the kaolin structure is comprised of a layer of silica tetrahedra (silicon ions, each of which is surrounded by four oxygen ions), and a layer of aluminum octahedra (aluminum ions, each of which is surrounded by 6 oxygen or hydroxyl ions). In other clay minerals other elements, such as magnesium, can substitute for aluminum in the octahedral sites. The combination of one tetrahedral layer and one octahedral layer is known as a 1:1 layer silicate. For simplicity it may be useful to describe this as a T-O structure. This structure is also found in the mineral serpentine.



Structure of the mineral kaolin



In a silica tetrahedron Si is surrounded by four oxygens. In an alumina octahedron Al is surrounded by six oxygen and hydroxyl ions.



The structure of illite is more complicated than that of kaolin. In this case each “sheet” within the structure is comprised of two tetrahedral layers (one “right side up” and the other “up-side down”). Sandwiched between these is an aluminum octahedral layer. Illite also has potassium ions situated at specific sites between the sheets. The combination of two tetrahedral layers surrounding one octahedral layer is known as a 2:1 layer silicate. We can

Structure of the mineral illite

describe this as a T-O-T structure. Some other T-O-T phyllosilicates include smectite, muscovite, biotite and chlorite.

The potassium cations (K^+) of illite are held in place by virtue of the fact that the upper and lower surface of each layer is “wall-to-wall” oxygen ions (O^{2-}). In other words these surfaces are negatively charged. These negatively charged surfaces are one of the fundamentally important features of all clay minerals, because such

Negative charges on clay mineral surfaces

surfaces are attractive to positively charged ions, such as heavy metals, or some organic pollutants. Not only do clays have these attractive surface areas, but they have very large surface areas. It is estimated that a cubic centimetre of clay has a reactive surface of around 2800 square metres, which is equivalent to the area of a football field! A centimetre-thick pad of paper includes about 100 sheets. A centimetre-thick layer of clay includes about 100,000 sheets. In other words a single sheet of paper is equivalent in thickness to approximately 1000 sheets of clay.

The cation exchange capacities of clay minerals and some other substances are listed in the following table.

Mineral/substance	CEC (meq/100 g)	pH dependance
kaolinite	3 to 15	strong
illite and chlorite	10 to 40	slight
smectite	80 to 150	negligible
zeolites	100 to 400	negligible
organic matter	100 to 500	strong
Fe and Mn oxy-hydroxides	100 to 740	strong
synthetic ion exchange resins	290 to 1020	slight

After D. Langmuir, (1997) (meq = milligram equivalents)

Cation exchange capacities of clays and other materials

The clay minerals, especially smectite, are strong adsorbers - much stronger than other minerals such as feldspar, but they aren't as good as things like organic matter and the hydrated oxides of iron and manganese, and don't come close to some synthetic ion-exchange materials.

The family of clay minerals known as the smectites³, which also have a T-O-T structure, are generally similar to illite, except that the inter-layer cations are different, and there are generally fewer of them; and also that other elements, such as iron and magnesium can substitute for aluminum in the octahedral layer. Sodium, calcium or magnesium are the common inter-layer cations in smectites, and in many cases more than one of these will be present. If the inter-layer cations are either magnesium or calcium up to two layers of water molecules can get in and push the sheets apart. If sodium is the inter-layer cation, even greater amounts of water can make its way in between the sheets, pushing them far apart - causing the mineral to swell dramatically. As a result of this property the smectites are known as swelling clays. This swelling of smectites can have very significant environmental geology implications. Buildings built on smectite-bearing soil can be damaged as the soil swells and shrinks in response to water levels. Water-saturated smectite-bearing rock is commonly very weak. Furthermore, because smectite-bearing rock can absorb so much water, these types of rocks can become overly heavy. The excess weight can contribute to slope instability.

Smectites

³ Some of the smectites are: *montmorillonite*, *beidellite* and *saponite*. Most are derived from the weathering (hydration) of mafic volcanic deposits. The term *bentonite* is used to describe a mixture of montmorillonite and beidellite. The term *Fuller's earth* is used for smectitic clays that have a high absorptive capacity. Molecular-level mixtures of smectite and illite are very common, in fact most illites have some smectitic layers, and most smectites have some illitic layers.

1.3 The atmosphere, oceans and climate

We are in the midst of a significant change in the focus of earth science - from an emphasis on studying minerals, rocks, resources and tectonics - towards the application of geological principals and techniques to an understanding of the interactions of the atmosphere and oceans with the solid earth, and the implications that these interactions have on our environment. This “new” direction for earth science is popularly known as “Earth System Science”.

Atmosphere,
oceans and
climate

Earth System Science is the study of both natural and anthropogenic processes that take place and have taken place within and between the atmosphere, oceans and solid earth. In fact a major role of those involved in Earth System Science is to examine and understand past natural variations in environment and climate in order to determine whether or not the changes that we have observed in the past century are related to anthropogenic (human caused) or natural processes.

Earth System
Science

Geologists are in an excellent position to study and understand these variations, for several reasons:

- geologists have a unique appreciation of the extent of the earth’s history and the pace of geological change,
- geologists have experience in the interpretation of the geological record to study past events and, and
- geologists are familiar with the chemistry and physics of materials of the solid earth, and their relationships with the atmosphere, the hydrosphere and the biosphere.

The atmosphere is described as being around 300 km thick, but this is not realistic because the upper part of

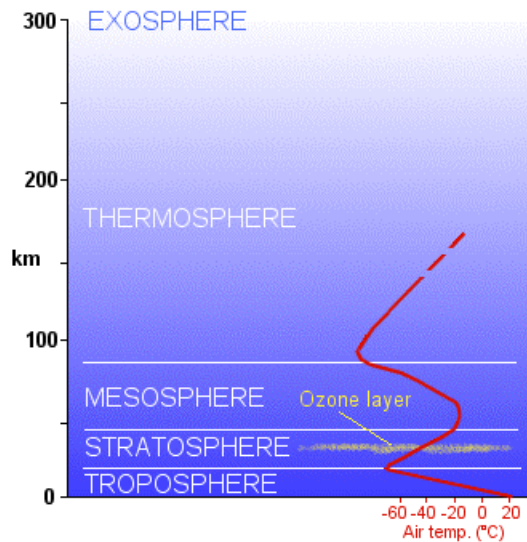
the atmosphere is exceedingly thin and it would be difficult to define where it ends. Over 90% of the mass of the atmosphere lies within 20 km of the earth’s surface.



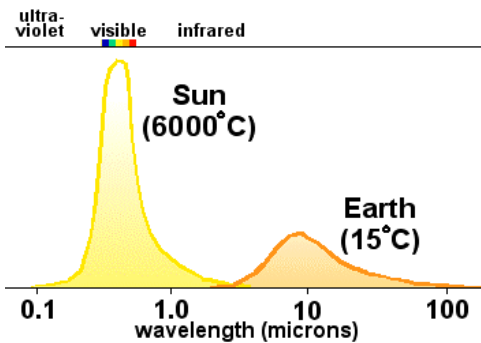
The atmosphere

The various “layers” of the atmosphere are shown on the diagram below. The two layers that are of greatest significance to us are the troposphere (which is where we live and breath) and the stratosphere.

Layers of the
atmosphere



The air temperature decreases with elevation in the troposphere⁴. The stratosphere has a higher level of ozone (O₃) than the troposphere, and ultraviolet light from the sun interacts with this ozone to create heat, such that the temperature gradient is reversed in the stratosphere. The boundary between the troposphere and the stratosphere is known as the tropopause.



The sun shines because it is hot. Its surface temperature is around 6000° C, and it emits light at wavelengths that are characteristic of a body at that temperature. Most of the light emitted by the sun is in the “visible” part of the spectrum (see diagram to the left), but there is also some in the ultraviolet and the infrared.

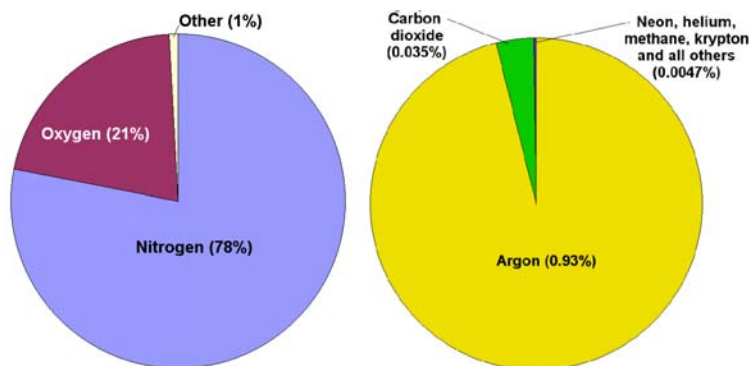
Solar radiation

The earth’s surface is heated primarily by the sun, and the average temperature at surface is around 15° C. The earth also emits light (much, much less than the sun), and all of that light is in the infrared part of the spectrum.

Terrestrial radiation

The atmosphere is made up of a number of different gases including about 71% nitrogen (as N₂), 28% oxygen (as O₂), 1% argon (Ar) and a few other inert gases. There are even smaller amounts of a number of important other gases, including ~ 0.04% carbon dioxide (CO₂), ~0.00017% methane (CH₄) plus even smaller amounts of nitrous oxide (N₂O) and the CFC gases (chloro-fluoro-carbons). These four gases, along with water, are known as **greenhouse gases**.

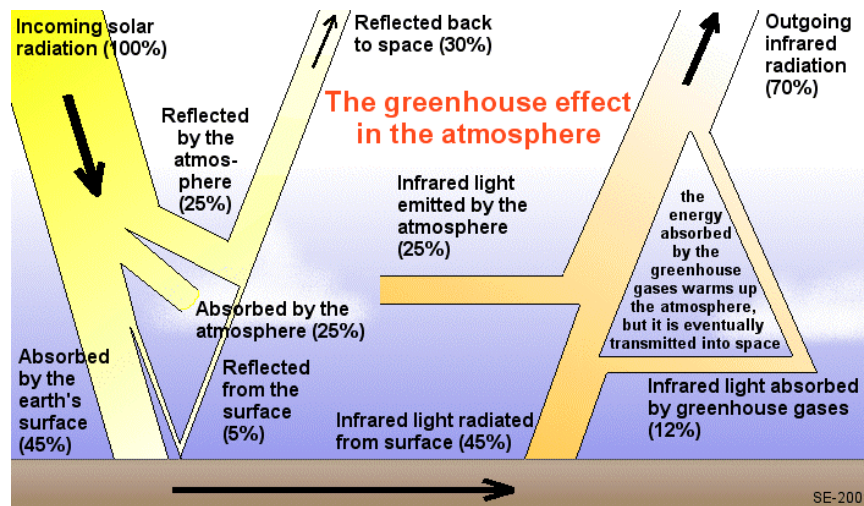
Composition of the atmosphere



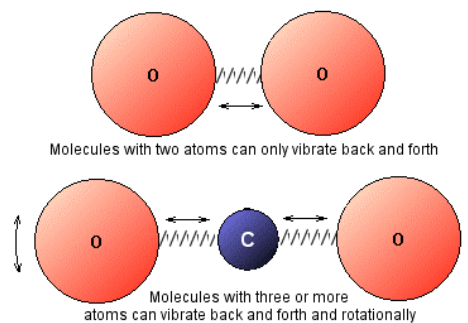
⁴ The rate of temperature decrease is known as the “lapse rate”, and that is about 6.5° C / km elevation.

As shown on the diagram below, approximately 50% of the sunlight (primarily visible) that enters our atmosphere makes it down to the earth's surface. The rest is either reflected directly back into space (mostly from the tops of clouds), or is absorbed into the atmosphere (mostly into clouds). The light that gets through leads to heating of the earth's surface, and that energy is then re-radiated back into space in the form of infrared light [see Figures 16.5 and 16.6 in Keller].

The greenhouse effect



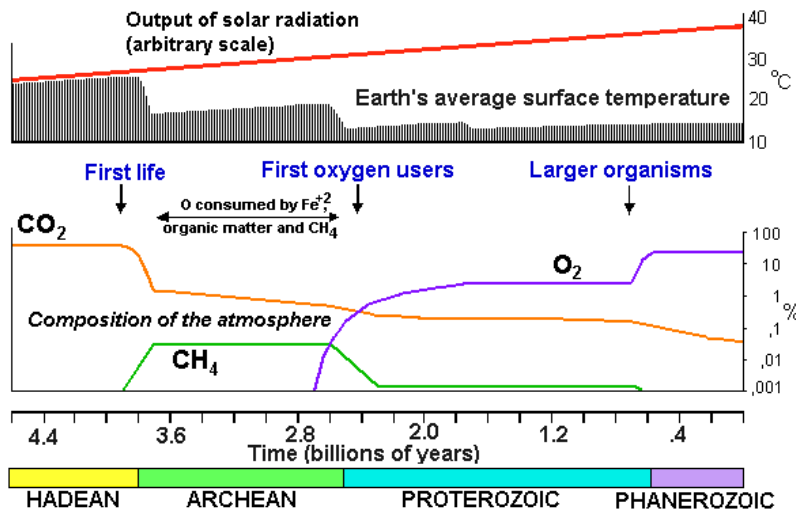
The sun's visible light does not interact with the gases of the atmosphere, but the re-radiated infrared light does interact with any gas that has more than 2 atoms. These types of gas molecules are able to vibrate in both translational (back and forth) and rotational senses, and their rotational vibrations are at the same wavelengths as some of the infrared light. When infrared light strikes one of these greenhouse gases it enhances the vibrations and causes the molecule to warm up. The greenhouse gases include water vapour, carbon dioxide, methane, nitrous oxide, ozone and the chlorofluorocarbons (CFC's). Excluding water vapour, carbon dioxide contributes about 60% of the greenhouse effect, followed by methane at 15% and the CFC's at 12% [Table 16.1 of Keller]. The greenhouse effect of water vapour is not clearly understood – but it has both warming and cooling effects because a lot of the water vapour ends up as water droplets in clouds and this reflects incoming sunlight.



Greenhouse gases

The greenhouse effect is not new. It has existed over geological time, and has systematically decreased in intensity as the luminosity of the sun has increased. (The sun's luminosity has increased by a factor of around 1.4 times over geological time.) For example, as shown on the figure below, it is likely that the earth's early atmosphere had a carbon dioxide content of greater than 10%. The level dropped to the 1850 minimum level of around 270 ppm (0.027%), but has since increased to over 350 ppm. The current methane level is around 1.8 ppm (1800 ppb), but prior to

oxygenation of the atmosphere there might have been over 100 ppm methane in the atmosphere.



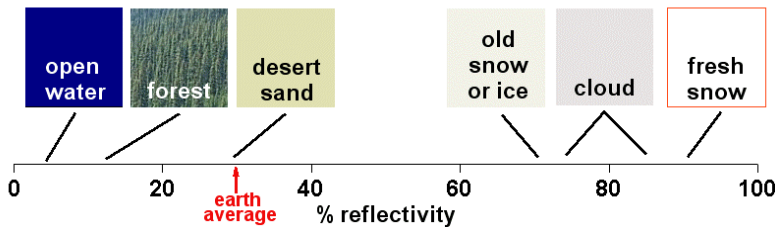
Summary of changes in atmospheric composition and temperature over geological time

from Lovelock, J., 1988, The Ages of Gaia

The albedo (reflectivity) of earth surfaces is important in controlling the amount of the sun's energy that is absorbed or reflected. Of the common surface types, ocean water has the lowest reflectivity, at around 5% (5% of the light is reflected, 95% is absorbed and converted to heat). Dense forest is close to 12%, and desert sand is around 30%. The most reflective surfaces are old snow and ice (~70%), cloud (70-90%) and fresh snow (~90%).

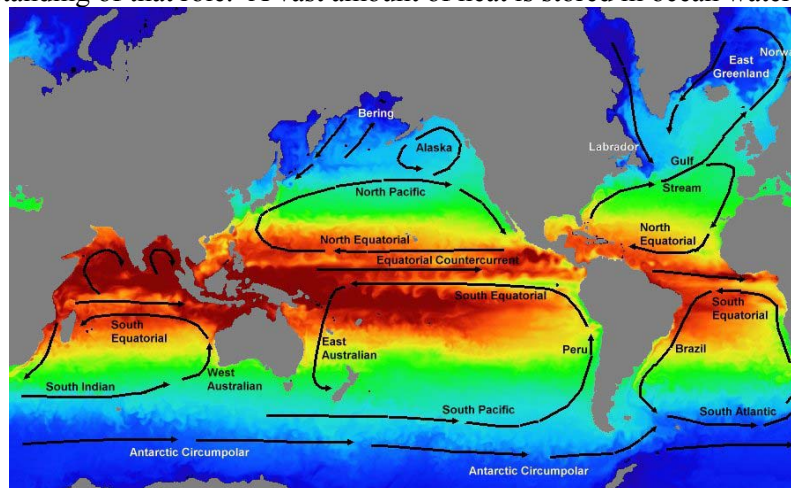
Earth's albedo

Albedo values for earth surfaces



The oceans

The oceans play a crucial role in climate and climate change, and it is important to have some understanding of that role. A vast amount of heat is stored in ocean water (partly because of its low albedo), and ocean currents are very important in the redistribution of heat on the earth - primarily the movement of warmth from the tropics to the poles, and vice versa.

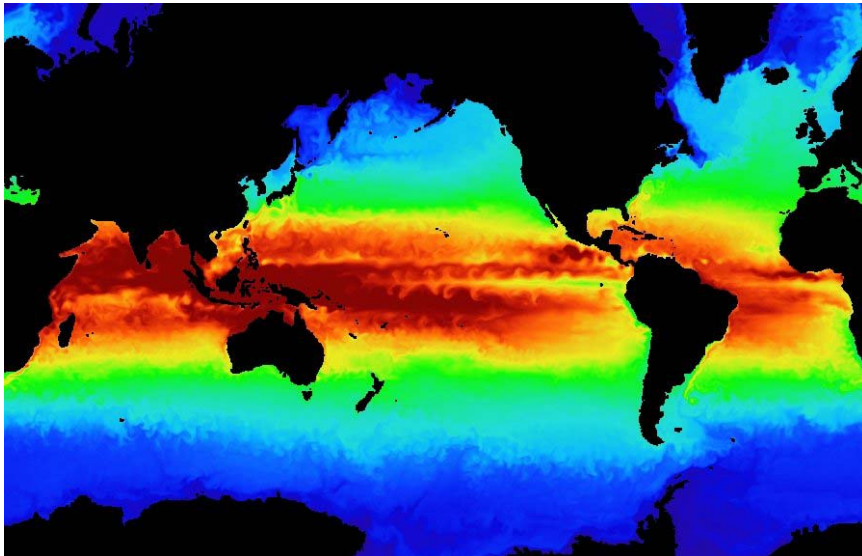


Major surface currents of the oceans

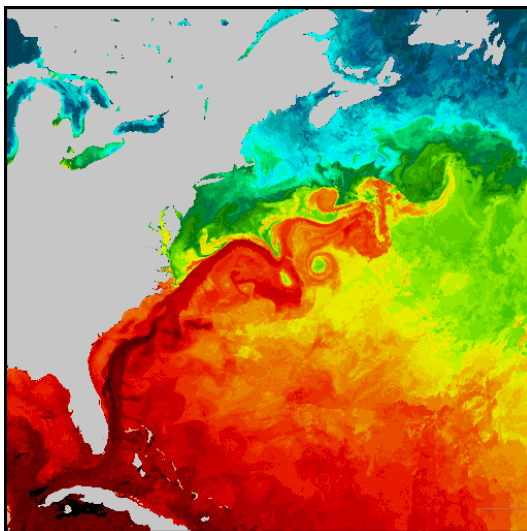
Ocean currents are ultimately driven by wind currents, although the resulting currents do not necessarily move in the same direction as the driving wind. South of the equator the Trade Winds blow towards the northwest, and north of the equator they blow towards the southwest. These winds create westerly-moving equatorial currents.

Controls on surface ocean currents

The ocean currents are also controlled by the coriolis effect, such that, for the most part, they rotate clockwise in the northern parts of the ocean basins, and counter-clockwise in the southern parts. The major currents have the general effect of moving warm tropical waters along the eastern margins of the continents towards the poles, and cold polar waters along the western margins of the continents towards the tropics. This redistribution of heat significantly impacts the earth's climate. Polar water has a temperature between 0 and 5° C (or even less than 0° for the southernmost Antarctic water) while tropical water is generally warmer than 25° C.



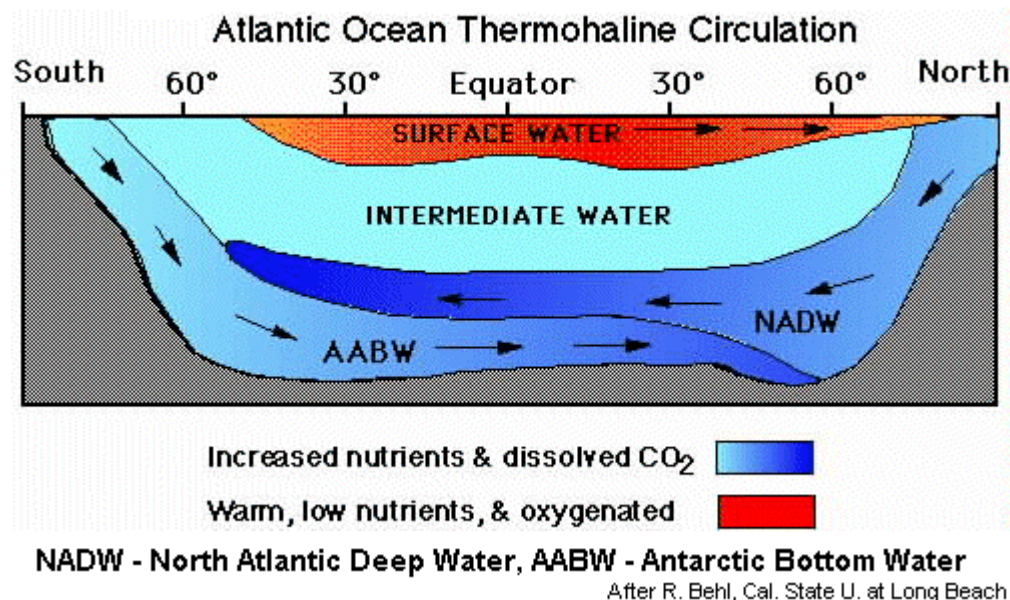
Global ocean sea-surface temperatures. Dark blue = cold (close to 0° C), Dark red = warm (close to 30° C)



North-south redistribution of the ocean's warmth can be seen clearly in the north Atlantic Ocean basin, where the Gulf Stream carries warm tropical water towards the northeast - keeping Europe relatively warm - while the Greenland and Labrador currents bring cold polar water to the south - giving Newfoundland a relatively cold climate, even though it is considerably farther south than Great Britain. Part of the Gulf Stream current can be seen on the figure to the left.

Temperature map of part of the Atlantic Ocean showing the Gulf Stream

The currents shown on the maps above represent only how the near-surface water (the upper few hundred metres) is behaving, but in fact much more exchange of heat takes place at depth in the oceans. The vertical currents in the Atlantic Ocean are shown on the map figure below.



Vertical component of ocean currents in the Atlantic Ocean

(There is more information on ocean currents at the following website:

<http://seis.natsci.csulb.edu/rbehl/300i-L09.htm>)

The distribution of these currents is controlled by what is known as thermohaline circulation because the density of seawater is a function of both its temperature and salinity. Tropical waters tend to be the most saline (because there is more evaporation taking place), while polar waters are by far the coldest. As shown on the figure above, the very cold water of the South Atlantic sinks to the ocean bottom to become Antarctic Bottom Water (AABW), while the slightly less cold water of the North Atlantic sinks to become North Atlantic Deep Water (NADW). The flow rates of these important systems have not been consistent over time. For example during the last glacial maximum (ca. 15,000 y ago) the flow of NADW may have been only 60% of what it is today (Thurman, 1997). Recent studies also suggest that the flow rate of the AABW may be as little as one-third of what it appears to have been over the past 800 years (Broecker et al., 1999). Variations of these types can (and are predicted to in the future) have significant implications for global climate.

Thermohaline circulation

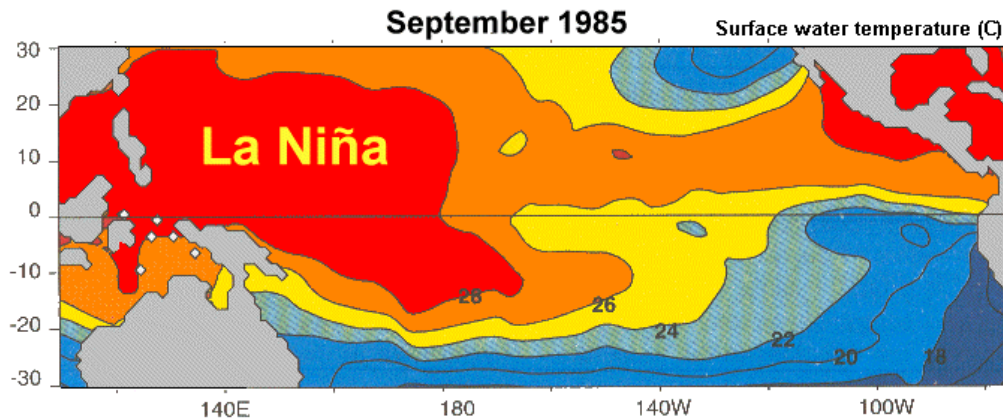
It is also clear that over longer time periods these patterns would necessarily change because of changes in the shapes and sizes of ocean basins related to plate tectonics. We will take another look at this concept when we discuss the causes and effects of glaciations.

Ocean current changes related to plate tectonics

An important example of how ocean circulation can affect climate is provided by the El Niño phenomenon of the equatorial part of the Pacific Ocean. Under “normal” conditions (known as La Niña) the Trade Winds blow strongly in a general westerly direction across the Pacific Ocean. These winds push a lot of the warm surface water (28 to 30° C) towards the west. In response, cold deeper water wells up to replace the

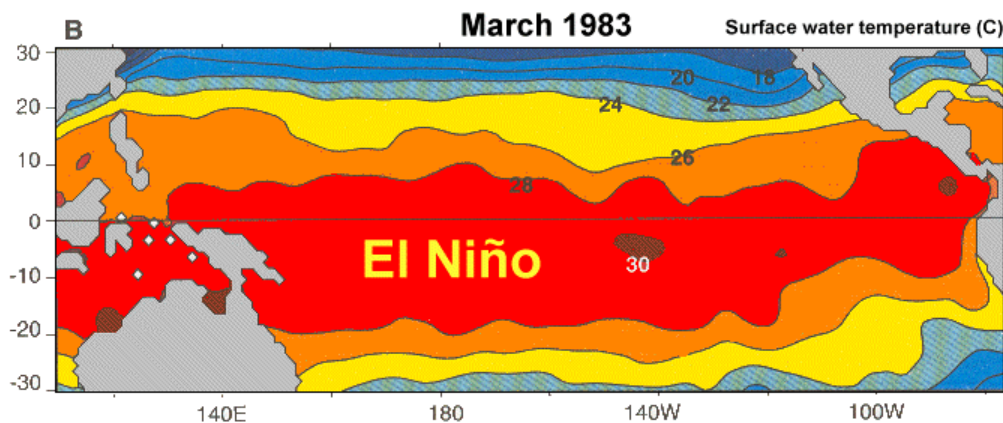
El Niño – southern oscillation

surface water in the eastern Pacific, producing surface-water temperatures of 24 to 25° C. The effect is self-sustaining because the higher temperature air of the eastern Pacific tends to rise and this strengthens the westerly Trade Winds.



Pacific Ocean sea-surface temperatures during a La Niña period

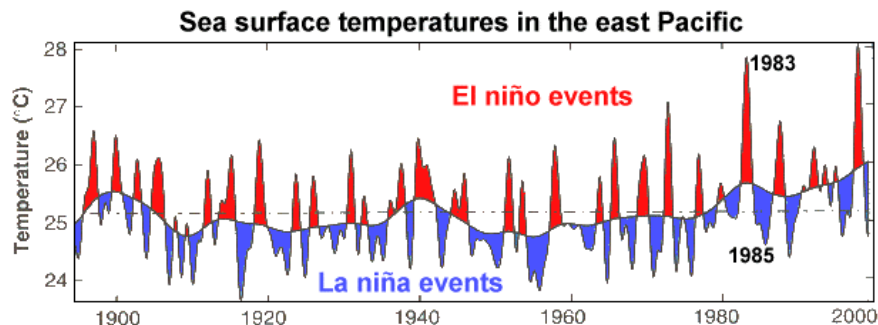
For reasons that are not well understood a La Niña condition can be overturned. A burst of easterly winds causes some of the warm water to move back towards the eastern Pacific, which in turn weakens the Trade Winds, and within a few months warm water extends right across the Pacific in an El Niño event. As described in the text [p. 461] an El Niño can have significant and dangerous impacts on the global climate - causing severe floods in some areas and killing droughts in others.



Pacific Ocean sea-surface temperatures during an El Niño period

For the past century La Niña has been the normal or more persistent of these oscillations. El Niño's have been spaced at intervals of 5 to 7 years and normally have lasted for no more than a year. A graph of sea-surface temperatures (SST) for the eastern Pacific is shown below. It is evident that El Niño's have become more extreme in the past two decades (higher peak temperatures), that the long-term average SST (the heavy line) has started to rise, and that the intervals between El Niño's have dropped to between 3 and 4 years. These changes are attributed to global warming.

Timing of El Niños



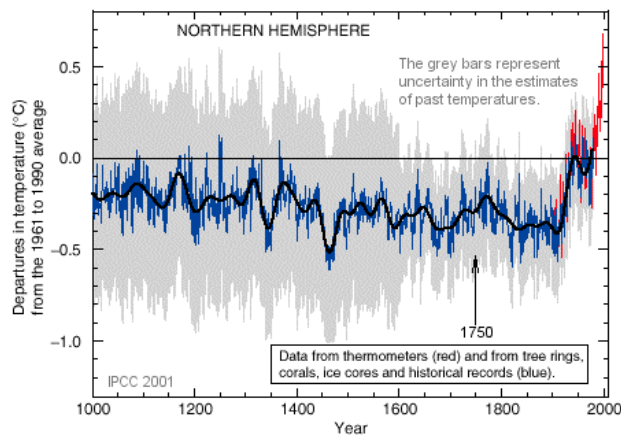
Variations in eastern Pacific Ocean sea surface temperatures since 1900

Climate change

Climate change

As noted above, the greenhouse effect is not new. What is new is that since the beginning of the period of industrialization we have contributed to significant changes in the composition of the atmosphere. Carbon dioxide levels are currently increasing at a rate of around 0.5% per year, and even this rate of increase is increasing (up from around 0.1% in the 1950's). Carbon dioxide is by far the most significant greenhouse gas, accounting for around 60% of the anthropogenic greenhouse effect. Methane levels are also increasing at around 0.5% per year, although this rate of increase appears to be diminishing - from a rate of over 1% in the 1980's (cf. Dlugokencky et al., 1998). In the 1970's and 1980's CFC levels were increasing at rates of between 4 and 6% per year (Cunnold et al., 1994, 1997). Since the United Nations Environment Programme **Montreal Protocol** of 1987, production of these gases has been dramatically reduced. The annual rate of increase of CFC-11 has leveled off, and rate of increase of CFC-12 is currently less than 1%. It is expected the levels of both types of CFC's will start to decrease in the next few years, and that they will return to pre-1980 levels by around 2050 (UNEP, 2000).

Variations in northern hemisphere temperatures over the past 1000 years are shown on the diagram to the right. Over the 900 years from 1000 to 1900 the average temperature dropped by a total of around 0.2° C. Over the past 100 years northern hemisphere temperatures have risen by close to 1.0° C.

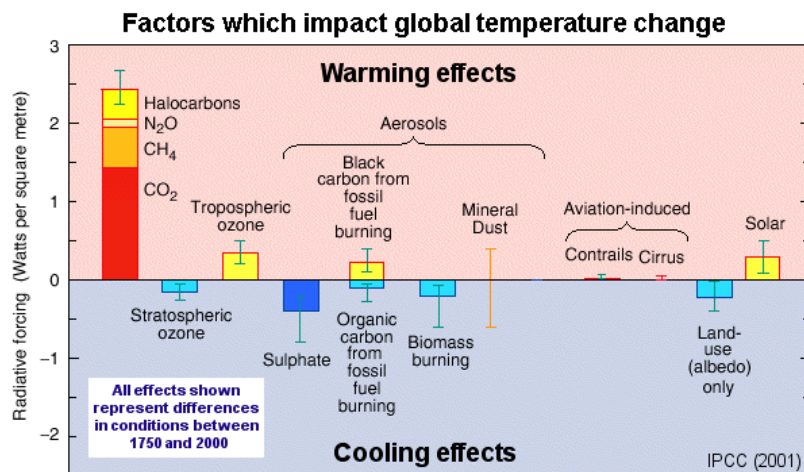


Northern hemisphere temperatures over the past 1000 years

The various factors that lead to climate change have been assessed by the UN-sponsored Intergovernmental Panel on Climate Change⁵, and are summarized on the diagram below. The changes in greenhouse gas levels are the major factor that has contributed to warming. Ozone in the troposphere (as opposed to the stratosphere) has a similar effect, but it is generally localized around cities, and is not a global effect. The most significant cooling effect comes from sulphate aerosols (drops and

⁵ IPCC: <http://www.ipcc.ch>

particles), some of which are natural (largely from volcanoes⁶) and some man-made.



Factors that provide positive and negative climate forcing

The IPCC predicts that warming over the next century will be much more significant than it was over the past century – probably at least 2 to 3 degrees and possibly 5 to 6 degrees C. This will have significant implications for plants and animals (many of which cannot adapt successfully to such rapid change) and will lead to serious sea-level rise. There will also be increases in storm frequency and intensity, more severe droughts in some drought-prone areas, and more serious floods in some flood-prone areas.

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⁶ One of the main volcanic gases – sulphur dioxide - combines with water to form sulphuric acid that exists as tiny droplets or solid particles in the atmosphere. These aerosols, and the ash particles, have the net affect of reflecting sunlight causing widespread cooling. For example, eruptions of Tambora (Indonesia) in 1816, Krakatau (Indonesia) in 1884 and Katmai (Alaska) in 1912 produced overall cooling of up to 0.5 C for relatively short periods (1 to 2 years) (Briffa et al., 1998). The June 1991 eruption of Mt. Pinatubo in Indonesia involved emission of 20 million tones of sulphur dioxide and 10 cubic kilometres of ash into the atmosphere.

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