EARTH AND PLANETARY SYSTEM

ORIGIN OF THE SOLAR SYSTEM

- After the ‘**Big Bang**’, the sequence of events:
  - ✓ At t = ~10^{-32} seconds, matter existed as a mixture of quarks (the fundamental building blocks of matter).
  - ✓ By t = 13.8 seconds, the universe cooled sufficiently for quarks to combine and form neutrons and protons, and then H, D, and He nuclei. The temperature was still too high for electrons to combine with the nuclei to form neutral atoms.
  - ✓ At = ~700,000 years, temperatures cooled sufficiently for electrons to attach to nuclei and form neutral atoms. Matter could then aggregate to form stars, galaxies, planets, etc.

- ***Solar nebular theory***: Formation of the sun & planets from a gaseous nebula - C.F. Von Weiszacker & Kuiper. The **nebular hypothesis** argues that diffuse, slowly rotating clouds of hydrogen and helium contract under the force of gravity. Contraction accelerates the rotation of the particles. Matter drifts to the center of the cloud and a proto-Sun may form. Compressed under its own weight, the proto-Sun becomes hotter and hotter until a temperature is reached where hydrogens fuse to form heliums and energy is released.

- Einstein showed that Energy is equal to the square of the speed of light times the mass. A small mass can produce a large amount of energy.
- In our solar system, the composition of the planets is a function of distance to the Sun. The inner planets (Mercury, Venus, Earth and Mars) are small and made up rocks and metals. Forming close to the Sun, the low density gasses boiled away and were not retained in quantity. The outer planets (Jupiter, Saturn, Uranus and Neptune (Pluto is thought to have been captured as a solid object) are sometimes referred to as the **gas bags** and are made up of ices and gases.

- The *inner* or **terrestrial planets**: Mercury, Venus, Earth and Mars are composed of largely silicates and metals while the *outer* or **Jovian planets**: Jupiter, Saturn, Uranus, Neptune and Pluto are composed chiefly of hydrogen, helium, methane, ammonia, and water. The inner four planets and moon have *densities* between 3.3 and 5.5 gm/cm3 while the outer planets have densities less than 2.0.

- **Asteroids –** small planetary bodies that revolve around the sun between the orbits of Mars and Jupiter. These are remnants of planetesimals (small planets) and other debris left over from the formation of the solar system and composed chiefly of silicates and iron-nickel. Meteorites are those asteroids which have landed on earth. But however some meteorites have come from moon and mars.

- **Meteorites –** are classified as *stones*, *stony irons* or *iron* depending on the relative contents of **silicates** and **iron**. *Stony meteorites* are most abundant and are further sub-divided into *chondrite* (silicate spherules, about 1mm in size) and *achondrites* (exhibit a wide variety of textures some which are indicative of crystallization from magma). Chondrites represent primitive planetary material from origin of the solar system and have not melted since their formation. On the other hand, irons, stony irons and achondrites are meteorites that have form by melting and crystallization of primitive planetary materials. Iron meteorites fragments of asteroid cores.

AGE OF THE EARTH AND SOLAR SYSTEM

- Oldest rocks are 3.8-4.0 Ga in age and they occur as small, isolated terrains in most continents which are known as older green stone belts.
- Isotopic dates from the moon 4.6Ga and age of the lunar crust being about 4.5 Ga.
- Isotopic dates from meteorites tell us the age of magmatic crystallization of the parent bodies from which the meteorites come. The oldest meteorites ages are about 4.6 Ga.
- Age of ancient lead ores on earth date from 4.6 to 4.4 Ga and record the time at which lead separated from other elements during formation of the earth's core.
- Age of the sun based on rate of energy loss of about 4.7 Ga.
- All the above lines of evidence suggested 4.6Ga (i.e. 4.6 billion years) for the earth.
- Age of the universe equal to 10-15 Ga (i.e. 10-15 billion years).
- Ga stands for Giga-annum

GEOLOGICAL CONCEPTS

- **Minerals** are the materials which make up the solid Earth and **rocks** are aggregates of minerals. Minerals are made up of **atoms** of the naturally occurring **chemical elements.**
- In general, there are three types of rocks:

  ✓ **Igneous** - form by the crystallization of liquid material - magma or lava
  ✓ **Sedimentary** - form in response to conditions at or near the surface of the Earth
  ✓ **Metamorphic** - form in response to changes in temperature and pressure.

- At some temperature and pressure the solid material may begin to melt. Rarely does the entire volume of rock melt. A mixture of melt and solids is produced by **partial melting**. The melt tends to rise (it is usually less dense than the material around it) and is called **magma**. If the magma is ejected onto the surface it is called **lava**.

INTERNAL STRUCTURE AND COMPOSITION OF THE EARTH

PLANETARY INTERIORS

Four independent study approaches are:

- Study of **ultramafic** rocks which are composed of mixtures of olivine, pyroxene, and garnets are representative of basaltic magma in the upper mantle. These rocks are brought to the earth surface by faulting and volcanic eruption.
- Measurement of **seismic-wave velocity** in rocks under high P-T conditions help to eliminate rocks end minerals that exhibit seismic wave velocities inconsistent with those observed in earth and to focus our attention on rocks that exhibit acceptable velocities.
- **Geochemical studies** at high pressure and temperature condition simulated in the lab help to synthesize minerals and rocks that may occur in the lower mantle and core.
- **Studies of meteorites** provide our only direct sample of planetary interiors.

INSIDE THE EARTH

- The size of the Earth -- about 12,750 kilometers (km) in diameter-was known by the ancient Greeks, but it was not until the turn of the 20th century that scientists determined that our planet is made up of three main layers: *crust, mantle,* and *core.*
- This layered structure can be compared to that of a boiled egg.
- The *crust,* the outermost layer, is rigid and very thin compared with the other two.
- Beneath the oceans, the crust varies little in thickness, generally extending only to about 5 km.
- The thickness of the crust beneath continents is much more variable but averages about 30 km; under large mountain ranges, such as the Alps or the Sierra Nevada, however, the base of the crust can be as deep as 100 km.
- Like the shell of an egg, the Earth's crust is brittle and can break.
- Below the crust is the mantle, a dense, hot layer of semi-solid rock approximately 2,900 km thick.
- The mantle, which contains more iron, magnesium, and calcium than the crust, is hotter and denser because temperature and pressure inside the Earth increase with depth. As a comparison, the mantle might be thought of as the white of a boiled egg.
- At the center of the Earth lies the core, which is nearly twice as dense as the mantle because its composition is metallic (iron-nickel alloy) rather than stony.
- Unlike the yolk of an egg, however, the Earth's core is actually made up of two distinct parts: a 2,200 km-thick liquid outer core and a 1,250 km-thick solid inner core.

- As the Earth rotates, the liquid outer core spins, creating the Earth's magnetic field.
- Not surprisingly, the Earth's internal structure influences plate tectonics.
- The upper part of the mantle is cooler and more rigid than the deep mantle; in many ways, it behaves like the overlying crust. Together they form a rigid layer of rock called the lithosphere (from lithos, Greek for stone).
- The lithosphere tends to be thinnest under the oceans and in volcanically active continental areas, such as the Western United States.
- Averaging at least 80 km in thickness over much of the Earth, the lithosphere has been broken up into the moving plates that contain the world's continents and oceans.
- Scientists believe that below the lithosphere is a relatively narrow, mobile zone in the mantle called the asthenosphere (from asthenes, Greek for weak). This zone is composed of hot, semi-solid material, which can soften and flow after being subjected to high temperature and pressure over geologic time.
- The rigid lithosphere is thought to "float" or move about on the slowly flowing asthenosphere.

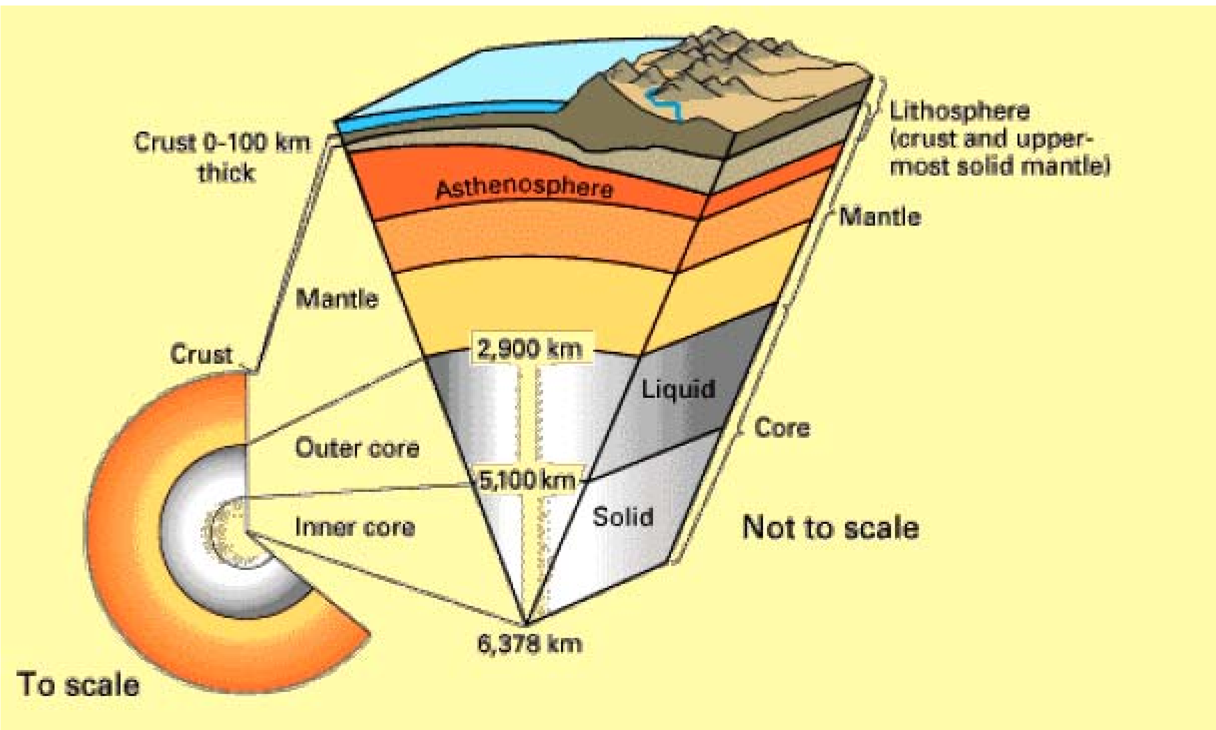

Figure showing the cutaway views of the internal structure of the Earth. This view drawn to scale demonstrates that the Earth‘s crust literally is only skin deep. Above right: A view not drawn to scale to show the Earth‘s three main layers (crust, mantle, and core) in more detail.

INTERNAL STRUCTURE OF THE EARTH BASED ON SEISMIC WAVE VELOCITIES

- Thickness of continental crust = 30-40 km
- Thickness of oceanic crust = 5 km
- Thickness of lithosphere (crust + upper part of the upper mantle) = 100 km
- Thickness of mantle = 2885 km
- Thickness of outer core = 2270 km
- Thickness of inner core = 1216 km

MINERAL ASSEMBLAGES IN THE EARTH'S MANTLE :

Depth (Km)	**Mineral Assemblages**
Upto 410 Km	Olivine, Pyroxenes, Garnet
410 Km	Olivine-Spinel phase transition (because of high pressure)
410-660 Km	Spinel, Pyroxenes, Garnet
660 Km	Spinel transforms into Mg-silicates with perovskite and periclase structures, pyroxenes, and garnet transforms into high-density phases
660-2900 Km	Mg-silicates with high-density crystal structures
> 2900 Km	metallic Fe-Ni (± sulphur and oxygen)

GENERAL CHARACTERISTICS OF SEISMIC WAVES IN THE EARTH'S INTERIOR

- P-waves (primary waves): characterized by particle motion parallel to direction of propagation. S-waves (secondary waves): also known as shear waves, characterized by particle motion normal to direction of propagation. Velocity of P-waves always greater than S-waves and S-waves cannot be transmitted through liquids.

- The mantle shows a sharp increase in velocity of seismic waves in comparison with those traveling in the crust and these velocities increase steadily with depth. The increase in seismic velocity suggests that the mantle is comprised of *eclogite* (garnet + pyroxene), *dunite* (olivine) and *peridotite* (olivine + pyroxene). The meteorites groups known as chrondites are similar in composition to the mantle.

- Abrupt changes in seismic wave velocities define the ***discontinuities*** which are 3 in number:

- ***Moho*** separates the crust from the mantle. The moho ranges in thickness from about 5 km in oceanic areas to over 40 km in oceanic areas. The material above the moho is called the crust which comprise of a upper granitic layer (SIAL) overlying a basaltic layer (SIMA). However the SIAL is absent in the oceanic crust. The crust along with the overlying sedimentary cover amounts to less than 0.5% of the radius of the earth. The moho represents a change from gabbroic rocks in the lower crust to ultra mafic rocks in the upper mantle.

- At 410km and at 660 km, ***secondary discontinuities*** occur in the mantle because of phase transition as defined above.

- The most striking discontinuity in the interior occurs at 2900 km and represents the core-mantle interface (***Guttenberg Discontinuity***).

- Another discontinuity at about 5200 km divides an outer core which is liquid (doesn't transmit S-waves) from an inner core which is solid, which is probably Fe-Ni, with low rigidity. The core comprises 16%by volume and 32% by mass of the total Earth. The density of the inner core is about 12.

- The zone lying between the discontinuities at 410km and at 660km is called the zone of transition or transitional zone. The transition zone is characterized by low seismic velocity (**LVZ**). The transition zone reflects partial melting. The region above is known as the *lithosphere*, behaves as a brittle solid and includes both the crust n part of the upper mantle. The LVZ is a zone of de-coupling that allows the plates to move. The region between the base of the lithosphere and the 660km discontinuity is called the *asthenosphere* and it deforms by plastic deformation or creep. The region between the 660km discontinuity and the core-mantle interface is known as the *mesosphere*, which is a part of the mantle i.e. strong yet passive in terms of deformation.

Structure of the Earth's interior based on seismic wave velocity distribution.

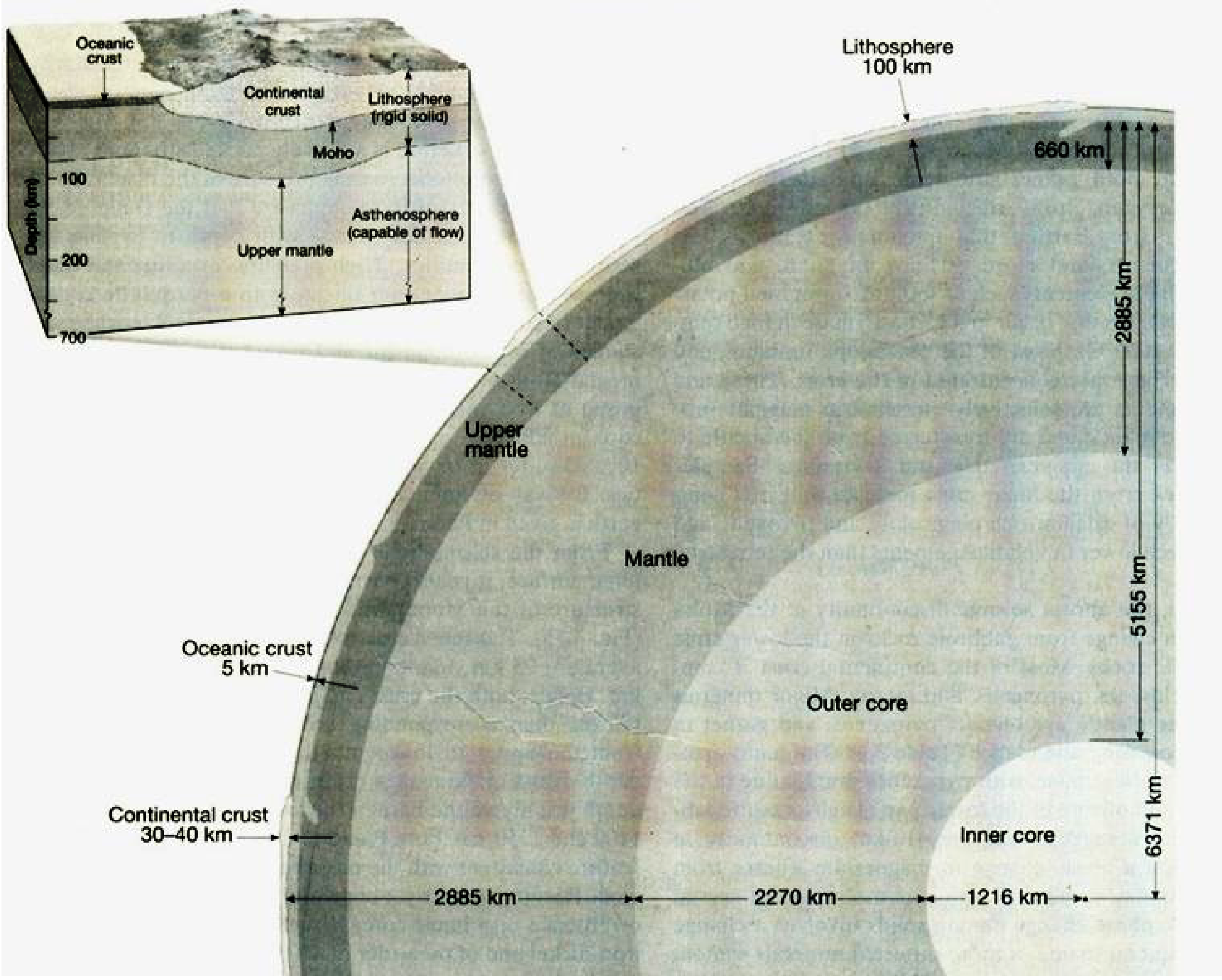

Figure showing the Structure of the Earth's interior based on seismic wave velocity distribution.

ELEMENTAL COMPOSITION OF THE EARTH (IN WEIGHT %)

Element	**Weight (%)**
Oxygen	31
Iron	30
Silicon	15
Magnesium	16
Calcium	1.5
Aluminium	1.3
Nickel	1.7
Sodium	0.9
Sulphur	2.0
Potassium	0.02
Titanium	0.09

- The earth is made up of 4 elements: oxygen, iron, silicon and magnesium.
- The moon differs from the earth in composition by comprising lesser metals, lesser volatile elements and more calcium, titanium and aluminium. Volatile elements are those that occur in a gaseous phase a temperature above 1000ºC: sodium, sulphur and potassium. Most of the uranium, potassium and thorium in the earth are concentrated in the crust.

CHARACTERISTICS OF THE EARLY CRUST

	Oceanic crust	**Continental crust**
First appearance	4.5 Ga	4.5 Ga
Where formed	Oceanic ridges	Subduction zones (trenches)
Composition	Komatiite-basalt	Tonalite-granodiorite
Lateral extent	Widespread	Local
How generated	partial melting of ultramafic rocks in the upper mantle	partial melting of wet mafic rocks in the descending lithosphere

- Three aggregation – break up super continent cycle
  - ✓ at 200 Ma.
  - ✓ at 600 Ma.
  - ✓ at 1500 Ma.

ANCIENT TECTONIC SETTING – ROCK ASSEMBLAGES

- **Arcs or collisional zones:** Ophiolites e.g. Alps Himalaya
- **Subduction zones: (Convergent plate boundaries)**
  - ✓ Accretionary prism – Melange (Ophiolites, deep sea sediments, arc volcanics and sediments)
  - ✓ Fore arc and intra arc basin – Greywackes (poor sorted clastic sediments) + volcanic tuffs
  - ✓ Volcanic and granitic rocks in arc system – Rhyolite (felsic volcanic rock, high in potassium, feldspar + quartz); Andesite (volcanic rock composition between rhyolite and basalt)
  - ✓ Metamorphic mineral assemblage – At shallow depth, blueschist facies (glaucophane)
- **Continental drift zone:**
  - ✓ Sediments – Arkose (feldspar rich), conglomerates, evaporates
  - ✓ Volcanic rocks – basalt and rhyolite
  - ✓ Deep igneous rock – granite
- **Cratons:**
  - ✓ Rifted continental margin (passive continental margin bounding opening ocean basin) e.g. Atlantic ocean
  - ✓ Platform areas
  - ✓ Continental margins of back arc basins
  - ✓ Marine quartz schist, shales and carbonate
- **Collisional mountain belts:**
  - ✓ Most distinctive collisional rock assemblage – suture assemblage (highly sheared melange separating foreign continental segments may include ophiolites, arc assemblages)
  - ✓ Foreland basin – basin on the descending plate – variety of sand stone and shales. These basins are shielded from the arc by a mountain belt.
  - ✓ Partial melting – granitic magma and ash flow tuffs (gaseous solid mixture at high temperature)

ATMOSPHERE AND GREENHOUSE EFFECT

ATMOSPHERE

INTRODUCTION

- Both the atmosphere and oceans were degassed from Earth's mantle chiefly in the first 50 Million years after planetary accretion began. Oxygen in the air, however, comes from photosynthesis, and its growth in the atmosphere parallels the expansion of algae in the Proterozoic and land plants in the Phanerozoic.

- The atmosphere comprises gaseous molecules that are continuously in motion. Because of the gravitational attraction of Earth, more molecules occur near the base of the atmosphere than at the top. The pressure of the atmosphere at any point is equal to the downward gravitational force of all the molecules in a vertical column above a given point, and the pressure is greatest at the base of the atmosphere and gradually decreases to near zero above 100 km height. Pressure is measured in units of bars or atmospheres, and the average pressure at the bottom of the atmosphere is one bar, which is approximately one atmosphere.

- Earth's present atmosphere is composed chiefly of nitrogen (78%) and oxygen (21%) with very small amounts of other gases, such as argon and carbon dioxide, and in this respect is unique among planetary atmospheres.(as shown in the table below)

Comparison Of Planetary Atmospheres

	Surface Temperature (^{o}C)	Surface Pressure (bars)	Principal Gases
Earth	-20 to 40	1	Nitrogen, oxygen
Venus	450 to 500	90	Carbon dioxide
Mars	-130 to 25	0.01	Carbon dioxide
Jupiter	-140	2	Hydrogen, helium
Saturn	-150	2	Hydrogen, helium
Uranus	-150	5	Hydrogen, methane
Neptune	-150	10	Hydrogen
Pluto	-200	0.005	Methane

- The concentrations of gases in Earth's atmosphere are controlled by several different processes. Oxygen, nitrogen, and carbon dioxide distributions are controlled by volcanic eruptions and by interactions between these gases and the solid Earth, the oceans, and living organisms. The distribution of minor gases, such as carbon monoxide, hydrogen, and ozone, is controlled primarily by reactions in the upper atmosphere caused by ultraviolet radiation from the Sun. Such reactions are known as photochemical reactions or photolysis, and Ultraviolet radiation is short-wavelength radiation that is lethal to many organisms.

- One important photochemical reaction is the fragmenting of oxygen molecules (O_2) into free oxygen atoms. These atoms are unstable and recombine to form new oxygen molecules or ozone, which is a molecule containing three oxygen atoms (O_3) as shown in the figure below. This reaction occurs today at heights of 30 to 60 km, with most ozone collecting in a relatively narrow band at about 25 km. Ozone is chemically unstable and breaks down to form molecular oxygen (plus a ***free*** oxygen atom that combines with another free oxygen atom to produce another oxygen molecule).
- Today, the production rate of ozone is approximately equal to it rate of loss, and thus the ozone band maintains a relatively constant thickness in the upper atmosphere. Ozone is an extremely important constituent in the atmosphere because it absorbs ultraviolet radiation from the Sun, and such radiation is lethal to many living organisms. Hence, ozone provides an effective shield which permits a large diversity of organisms to survive on our planet.
- Human-made pollutants introduced into our atmosphere in the last 50 years are reacting with ozone and beginning to destroy the ozone layer. A large hole in the ozone layer has developed over the South pole from such reactions. If atmospheric pollution continues, the ozone layer could be completely destroyed and many life forms on Earth could be threatened with extinction.

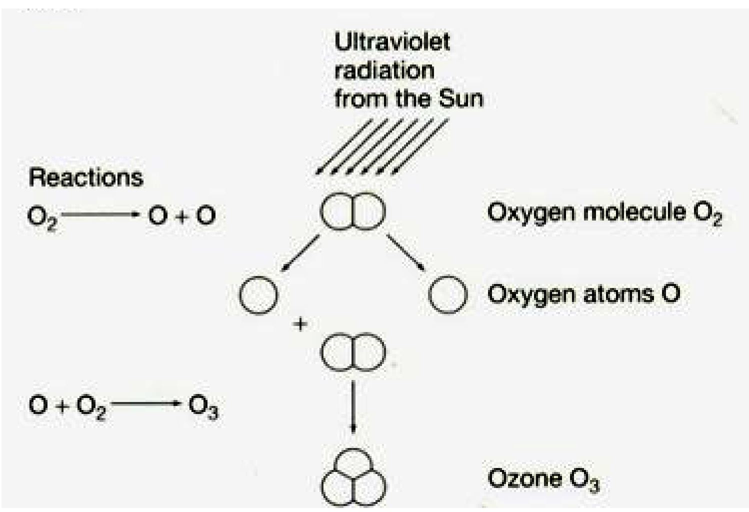

Figure showing the production of ozone in the upper atmosphere from photolysis caused by ultraviolet radiation from the Sun.

THE ORIGIN OF THE ATMOSPHERE

- Two sources are generally considered for Earth's atmosphere: left-over gases from planetary accretion, or degassing of Earth. The simplest origin is that the atmosphere consists of gases that were left over from the gaseous nebula after accretion of Earth. However, if our atmosphere were of this origin, it should be rich in hydrogen, helium, methane, and related components. Because this is not the case, our atmosphere is often referred to as a **secondary atmosphere**, implying that it was acquired after planetary accretion. Whether or not a primary atmosphere composed of left-over gases ever existed is a subject of some debate among specialists.

- The fact that volcanic eruptions release tremendous amounts of gas into the atmosphere indicates that Earth is losing gases from its interior by a process known as degassing. The abundance of volcanic rocks in the geologic record strongly suggests that large quantities of volcanic gas have entered the atmosphere in the geologic past and that the atmosphere and oceans may have formed by this process.
- Another line of evidence supporting a secondary origin for the atmosphere is the large amount of argon- 40 that it contains compared to the Sun. It is generally assumed that because the Sun and Earth condensed from the same gaseous nebula, they should have the same proportions of isotopes of various elements unless changed by later radioactive decay. Argon-40 is produced by the radioactive decay of potassium-40 in the solid Earth, and because it is a gas, it readily finds its way to Earth's surface and enters the atmosphere. Compared to the Sun and other planetary atmospheres, our atmosphere is rich in argon-40. This supports the idea that it was produced by degassing of Earth.

THE POSSIBILITY OF A PRIMARY ATMOSPHERE

- One line of evidence supporting the existence of an early primitive atmosphere is that volatile elements should collect around accreting planets during their late stages of formation. By analogy with the composition of the Sun, the atmospheres of the outer planets, and the composition of volatile-rich meteorites, an early atmosphere should he rich in such gases as hydrogen, helium, methane, and ammonia, and thus be a reducing atmosphere. A **reducing atmosphere** is one in which free oxygen or highly oxidized gases such as carbon dioxide cannot exist. Another important feature of an early reducing atmosphere is the possibility of producing organic compounds that may combine to form living organisms. There are arguments against the existence of an early reducing atmosphere, however.
- Although with present data we cannot prove or disprove that a reducing atmosphere existed prior to 4 Ga, if it did exist, it must have been lost before this time. If life originated in such an atmosphere, it is unlikely that it could have survived the loss of the atmosphere, and for this reason it appears that life did not form until the secondary atmosphere made its appearance. Because undeniable evidence of unicellular organisms occurs in rocks 3.6 Gy old, it is likely that a secondary atmosphere began to accumulate before this time.
- If an early reducing atmosphere existed, how was it lost? One idea is that it was blown away by a tremendous solar wind coming from the early Sun. Many star similar to our Sun evolve through a stage known as a T-Tauri stage, during which time large amounts of energy are expelled as a wind of high-energy particles that could easily blow volatile elements out of the inner part of the Solar System. If our Sun went through a T-Tauri stage, it would have

been soon after planetary accretion (by 4.6 Ga), and such an event should have removed any early atmosphere around Earth. Alternatively, an early atmosphere could have been lost during formation of the Moon. If the Moon formed by collision of a Mars-sized body with Earth, as most data favour, any primary atmosphere would probably have been lost during the collision.

THE SECONDARY ATMOSPHERE

- Degassing of Earth occurs directly by volcanism and by the weathering of igneous rocks at the surface, which liberates gases such as water vapor, and carbon dioxide. Of these two mechanisms, volcanic degassing has been shown to be by far the most important in atmospheric growth.
- To test further the degassing model for our atmosphere's origin, it is instructive to compare the composition of gases emitted by volcanoes with those in the atmosphere. To do this one must consider not only the atmosphere, but also the hydrosphere (oceans, lakes, rivers. etc.), the biosphere (living organisms), and those volatile elements tied up in sediments (such as carbon dioxide in limestones). This is because many of the volatiles degassed from Earth reside in these near-surface volatile reservoirs.
- The distributions are calculated on an oxygen-free basis because the source of oxygen in the atmosphere is clearly not from volcanic activity. Also, is the striking similarity between gases in near-surface reservoirs and volcanic gases. Water is the most abundant volatile in both categories, and carbon dioxide is the only other volatile of any consequence. Such a similarity, strongly suggests that Earth's near-surface volatiles have been derived from degassing of the mantle by volcanic activity.
- Two models for the composition of the early secondary atmosphere have been proposed, depending on whether or not metallic iron existed in the mantle when the secondary atmosphere began to accumulate from degassing (shown in the table).

Models For The Composition Of The Early Secondary Atmosphere

Metallic Iron Absent in Mantle	**Metallic Iron Present in Mantle**
Principal Gases : Carbon dioxide (CO_2) Water vapor (H_2O) Nitrogen (N_2)	Principal Gases: Hydrogen (H_2) Carbon monoxide (CO) Methane (CH_4)
Minor Gases: Hydrogen (H_2) Hydrogen chloride (HCl) Sulfur dioxide (SO_2) Nitrogen (N_2)	Minor Gases: Carbon dioxide (CO_2) Water vapor (H_2O) Hydrogen sulfide (H_2S)

- If metallic iron was present in the mantle, chemical reactions with the iron would produce gases rich in hydrogen, carbon monoxide, and methane. On the other hand, if metallic iron was absent, reactions with mantle silicates would produce gases rich in carbon dioxide, water vapor, and nitrogen.
- In terms of our current understanding of the formation of Earth’s iron core, it is difficult to imagine any model for the early Earth in which at least some metallic iron is not present in the mantle. However, because it is likely that the core formed in a relatively short period of time and that core formation effectively extracted iron from the mantle, the atmosphere may have changed rapidly in composition during the first 100 Million years of Earth history.
- Certainly by 4 Ga metallic iron would be cleaned out of the mantle and water, carbon dioxide, and nitrogen should be the most important gases entering the atmosphere.
- Several Factors support the existence of a terrestrial atmosphere rich in water and carbon dioxide by 4 Ga. For instance, organic compounds, a necessary prerequisite for living organisms, have been produced in the laboratory from such gases. Also, analyses of microscopic inclusions of gases in fragments of the mantle, which are brought to the surface during volcanic eruptions, indicate that carbon dioxide is the principal gas in the mantle.

THE GROWTH OF OXYGEN

- Earth is the only planet with free oxygen in the atmosphere and thus the only planet capable of sustaining higher forms of life. In the atmosphere today, oxygen is produced by two processes: photosynthesis and photolysis of water (as shown in the figure).
- During photosynthesis, algae and plants combine carbon dioxide and water to produce carbohydrates and oxygen, and the oxygen is liberated into the atmosphere. Much of the carbohydrate is converted back to carbon dioxide and water by respiration. During respiration, organisms absorb oxygen from the atmosphere and give off water and carbon dioxide. Decay of organic material, both in aquatic and terrestrial environments, also uses oxygen and releases carbon dioxide and water. Some oxygen is also extracted from the atmosphere by oxidation of surface rocks during weathering (for instance, in producing red iron oxides in soils) and by the oxidation of volcanic gases. Oxidation, however, is very minor compared to losses by respiration and decay.
- Today, the combined rate of carbon dioxide and water release by respiration, decay, and oxidation approximately equals the consumption rate of these gases by photosynthesis.
- A very minor amount of oxygen is also produced by photochemical reactions with ultraviolet rays in the upper atmosphere (as shown in the figure below).

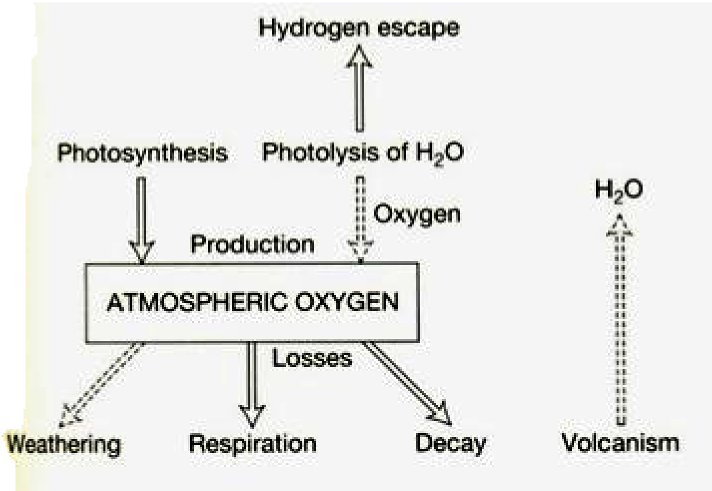

Figure showing the Major processes controlling production and losses of oxygen in the modern atmosphere. Solid lines indicate major controls and dashed lines minor controls.

OXYGEN CONTROLS IN THE PRIMITIVE ATMOSPHERE

- Prior to the appearance of photosynthetic microorganisms in the geologic record, and even for a considerable time thereafter, photosynthesis was not an important process in controlling oxygen levels in the atmosphere.
- Hence, in the primitive atmosphere, the oxygen content was controlled by the rate of photolysis of water; hydrogen loss into space; oxidation of volcanic gases; oxidation of iron in seawater, producing banded iron formation; and, after continents emerged above sea level, the weathering rate at the surface (as shown in the figure below). The rate at which water is supplied by volcanic eruptions is also important because volcanism is the main source of water available for photolysis.
- By 3.5 Ga, early photosynthesizing organisms may also have contributed oxygen to the primitive atmosphere. As photosynthesis became more widespread, recombination of hydrogen and oxygen to form water could not keep pace with oxygen input, and the oxygen level in the atmosphere continued to increase. This assumes that the rate of weathering did not increase with time, an assumption which seems to be supported by geologic data.

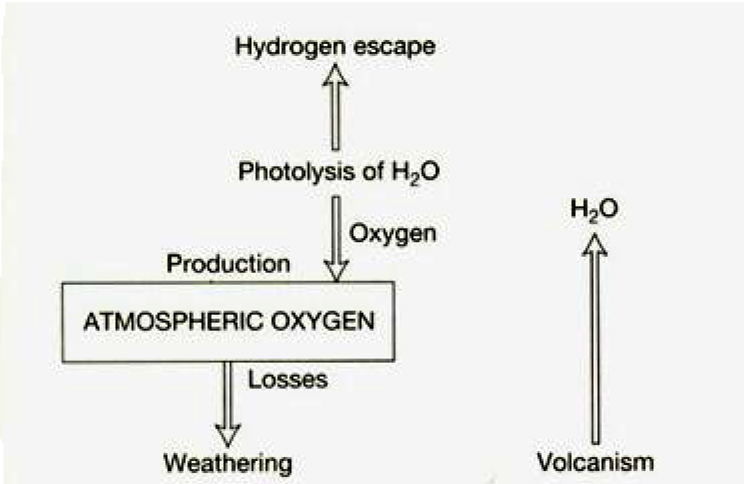

Figure showing the Major processes controlling production and losses of oxygen in the Early Archean atmosphere. Any arrows entering the boxes lead to increased oxygen and arrows leaving the boxes lead to decrease oxygen. Photosynthesis did not become significant until the middle Paleozoic.

GEOLOGIC INDICATORS OF ATMOSPHERIC OXYGEN LEVELS

- BANDED IRON FORMATION :

  - Banded iron formations, or BIFs, are sedimentary rocks, typically thin bedded or laminated with more than 15% iron.
  - BIFs are thought to have been chemically precipitated on the sea bottom.
  - Dark bands are principally magnetite or hematite, and the light bands are fine-grained, chemically deposited quartz.
  - Although most abundant in the Late Archean and Early Proterozoic, BIF occurs in rocks as old as 3.8 Ga and as young as 0.8 Ga.
  - Most investigators consider that BIF was deposited in large basins on cratons in shallow marine environments.
  - Many BIFs are characterized by thin, varvelike laminations that can be correlated over hundreds of kilometers.

- REDBEDS, SULFATES AND URANINITE :

  - Redbeds are sandstones and shales with red iron oxide cements. They generally form in stream environments, and the red cements are the result of oxidation of the iron soon after deposition.

  - Redbeds require significant amounts of oxygen in the atmosphere. The fact that redbeds do not appear in the geologic record until about 2.4 Ga and are not abundant until after 1 Ga suggests that oxygen levels were very low in the atmosphere during the Archean.

- Deposition of sulfates, primarily gypsum and anhydrite, requires free oxygen in the ocean and atmosphere. Although evidence of gypsum deposition is found in some of the oldest known sedimentary rock (~ 3.6 Ga), sedimentary sulfates do not become important in the geologic record until after about 2 Ga, thus supporting rapid growth of oxygen in the atmosphere, beginning in the Early Proterozoic.
- The occurrence of detrital uraninite (a uranium oxide mineral) in Late Archean to Early Proterozoic sedimentary rocks is well documented, and the best known examples are those in the Witwatersrand Basin in South Africa (~2.8 Ga) and those in the Blind River—Elliot Lake area in Canada (~2.3 Ga).
- No significant occurrences, however, are known to be younger than Mid-Proterozoic. Uraninite is unstable under oxidizing conditions and is rapidly dissolved by streams. The restriction of major sedimentary uraninite deposits to greater than 2.3 Gy again favors very low oxygen levels in the atmosphere prior to this time.

- PALEOSOLS :

  - Paleosols are ancient weathering profiles or soils that contain information about atmospheric composition.
  - For instance, today under present weathering conditions, the iron in soils developed on both granites and basalts is fully oxidized to hematite and related minerals.
  - In contrast, in paleosols older than 2 Ga, iron is fully oxidized only in soils that developed on granites, and a large amount of the iron in basaltic soils is not oxidized.
  - This is because the oxygen level of the atmosphere at that time was high enough to oxidize all iron released from granite weathering but not all of that released during basalt weathering.

- BIOLOGICAL INDICATORS :

  - The Precambrian fossil record also provides clues to the growth of atmospheric oxygen. Archean and Early Proterotoic cells were entirely primitive unicellular organisms, the earliest forms of which appear to have evolved in oxygen-free environments.