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The Earth as a Distant Planet
A Rosetta Stone for the Search of Earth-Like Worlds
Vázquez, M.; Pallé, E.; Montañés Rodríguez, P.
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Chapter 2: Earth in Time

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Chapter 2
The Earth in Time

The Earth is now characterized by a set of physical parameters and features observed on its surface. However, the Earth in its present state represents just one stage in an evolutionary process starting from the moment the planet formed to the present day. In our future search for Earth-like planets, we may encounter a planet at any stage in its evolution, and therefore, we can feasibly use the history of the Earth as a guide to characterize these planets. In this chapter we study the different components of the Earth system over time, starting with an introduction to the Earth as it is at present.

Most of the observed changes in the surface of the Earth have come about as a consequence of the dissipation of the energy stored in its interior. Variations in the gaseous envelope, the atmosphere, are mainly driven by changes in the solar output and in the concentration of greenhouse gases, together with a strong internal variability. Thus, we can divide all the changes suffered by our planet into two main classes: abiotic (solar and geological) and biotic.

A holistic view of our planet is based on the assumption that all its properties cannot be determined only by the sum of its components. This constitutes the core of the Earth System Sciences (Kump et al. 2004; Steffen et al. 2005).

We can consider our planet as constituted by four vast reservoirs of material with flows of matter and energy between them: atmosphere, hydrosphere, biosphere and geosphere (Fig. 2.1). The entire system evolves as a result of positive and negative feedbacks between constituent parts.

For Schellnhuber (1999), the Earth System, E, can be represented by the equation

$$\mathrm{E} = (\mathrm{N}, \mathrm{H}),$$

where N = (a,b,c,..) is the ecosphere and consists in a set of linked planetary sub-spheres: *a* (atmosphere), b (biosphere), *c* (cryosphere), and so on; H = (A,S) embraces the antroposphere A, and the component S reflects the emergence of a global subject, manifested, for instance, by adopting international laws for climate protection.

Table 2.1 and Fig. 2.2 show the main periods of the Earth history and the main geological and biological events, respectively, which have taken place throughout the evolution of our planet.

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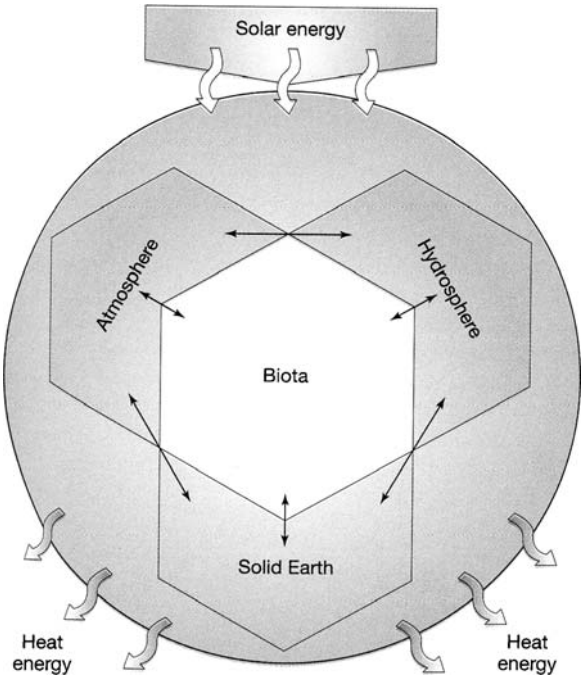

Fig. 2.1 Diagram of the Earth system, showing interactions among the components. From R.W. Christopherson, *Geosystems: an introduction to physical geography, 1997*. Copyright: Prentice-Hall

Table 2.1 Geologic periods

Precambrian	Hadean (4500–3800)		Moon formation
			Origin of the oceans
	Archaean (3800–2500)		Origin of life
	Proterozoic (2500–550)		Rise in atmospheric oxygen
			First cells with nucleus
Phanerozoic	Paleozoic	Cambrian (550–490)	Complex Life
		Ordovician (490–443)	
		Silurian (443–417)	
		Devonian (417–354)	
		Carboniferous (354–290)	
		Permian (290–248)	Massive extinction
	Mesozoic	Triassic (248–206)	
		Jurassic (206–144)	
		Cretaceous (144–65)	Massive extinction
	Cenozoic	Tertiary (65–1.8)	
		Quaternary (1.8–Today)	Humans

The time in brackets is in Ma

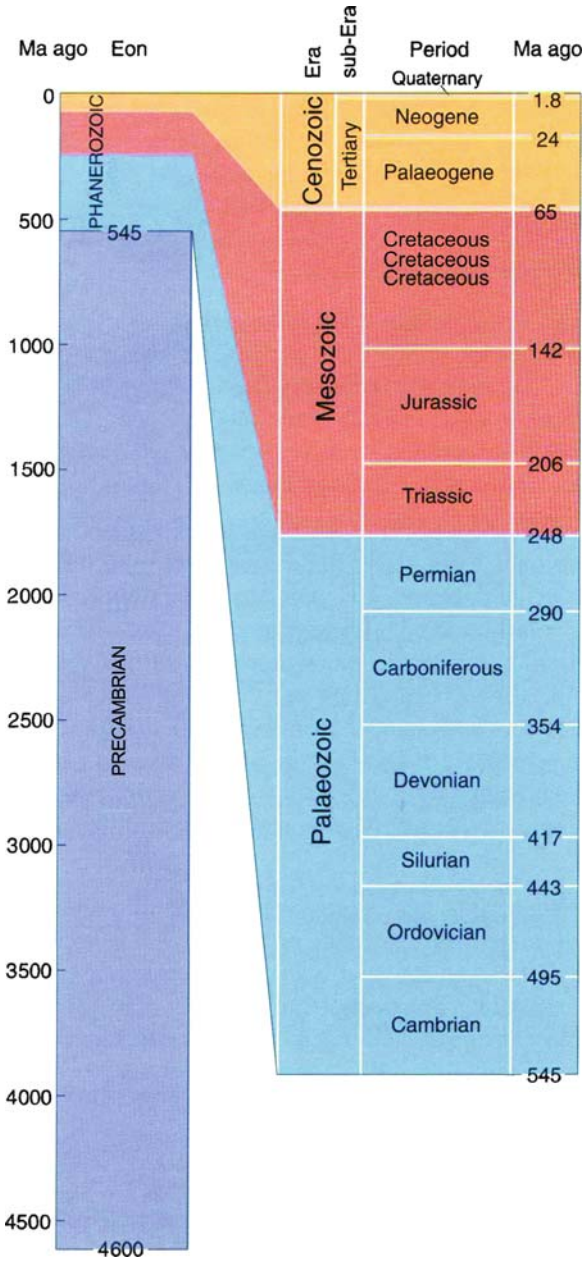

Fig. 2.2 Main geological periods

Different textbooks and monographs have handled the topic of this chapter. We can indicate only some of them (Stanley 1992, 1999; Lunine 1999; Lane 2002; Anguita 2002; Knoll 2003 and Zahnle et al. 2007).[1]

2.1 The Earth at the Present Time

The mass of our planet is approximately 5.98×10^{24} kg. It is composed mostly of iron (32.1%), oxygen (30.1%), silicon (15.1%), magnesium (13.9%), sulphur (2.9%), nickel (1.8%), calcium (1.5%) and aluminium (1.4%), with the remaining 1.2% consisting of trace amounts of other elements.

To understand the past and to predict the future, it is essential to know what is the present structure of our 4.6 Ga old planet. We follow a classical approach describing the different layers (Fig. 2.3) and the processes taking place there. Then we begin our time travel back to its past and into its future.

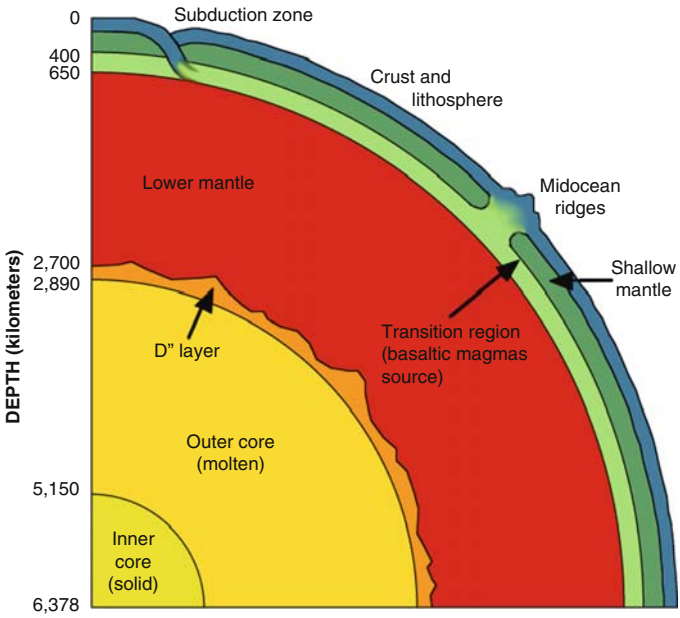

Fig. 2.3 Scheme of the different layers of the Earth, from the core to the surface

[1] See also the 2008 report of the National Research Council on *Origin and Evolution of Earth: Research Questions for a changing Planet*.

2.1.1 The Interior

The interior of the Earth is chemically divided into layers as a result of its molten state, early in its formation. Together with theoretical work, the main sources of information about this region comes from the recordings of free oscillations excited by large earthquakes.

The simplest model of the Earth is based on the assumption of an average density, $\overline{\rho} = 5.52\,\mathrm{g\,cm^{-3}}$. It is obvious that this value is higher than the average density (2.7–3.3) of the rocks of the Earth's surface. This clearly points to a concentration of mass near the centre of the Earth. Assuming that the planet is in hydrostatic equilibrium and is spherically symmetrical

$$\mathrm{dP} = -\rho \mathrm{g}\ \mathrm{dr}.$$

The gravity is given by

$$\mathrm{g} = \frac{\mathrm{Gm}}{\mathrm{r}^2} = \frac{4\pi \mathrm{G}}{\mathrm{r}^2} \int_0^{\mathrm{r}} \rho \mathrm{r}^2 \mathrm{dr},$$

where G is the gravitational constant and $m = \frac{4\pi}{3} r^3 \rho$ is the mass enclosed within the sphere of radius r and density ρ. Hence,

$$\frac{\mathrm{dP}}{\mathrm{dr}} = -\frac{4\pi \mathrm{G}\rho}{\mathrm{r}^2} \int_0^{\mathrm{r}} \rho \mathrm{r}^2\ \mathrm{dr}.$$

Now, by splitting the density variation in two parts

$$\frac{\mathrm{d}\rho}{\mathrm{dr}} = \frac{\mathrm{d}\rho}{\mathrm{dp}} \frac{\mathrm{dP}}{\mathrm{dr}},$$

we obtain

$$\frac{\mathrm{d}\rho}{\mathrm{dp}} = \frac{\rho}{\mathrm{K}},$$

where K is the compressibility module. Therefore,

$$\frac{\mathrm{d}\rho}{\mathrm{dr}} = -\frac{\rho}{\mathrm{K}} \frac{\mathrm{Gm}\rho}{\mathrm{r}^2}.$$

From the theory of propagation of waves in the interior of the Earth, we know that

$$\frac{\mathrm{K}}{\rho} = \alpha^2 - \frac{4}{3}\beta^2,$$

where α and β are the velocities of the P and S waves,[2] respectively, given by

$$\alpha = \left[\left(K + \frac{4}{3}\mu\right) \Big/ \rho\right]^{1/2}$$
$$\beta = (K/\rho)^{1/2},$$

where ρ is the density and μ is the rigidity modulus

$$\longrightarrow \frac{\mathrm{d}\rho}{\mathrm{dr}} = -\frac{\mathrm{Gm}\rho}{\mathrm{r}^2\left(\alpha^2 - \frac{4}{3}\beta^2\right)}$$

Moreover, we must also know the equation of state of the material in the interior that establishes a relation of density with temperature and pressure at all the depths.

The Preliminary Earth Reference Model constitutes a good framework to study the processes taking place in the Earth's Interior (Dziewonski and Anderson 1981). Figure 2.4 shows the variations with the radius of the most relevant parameters. For monographs on the Earth's internal structure see Zharkov (1986) and Poirier (1991).

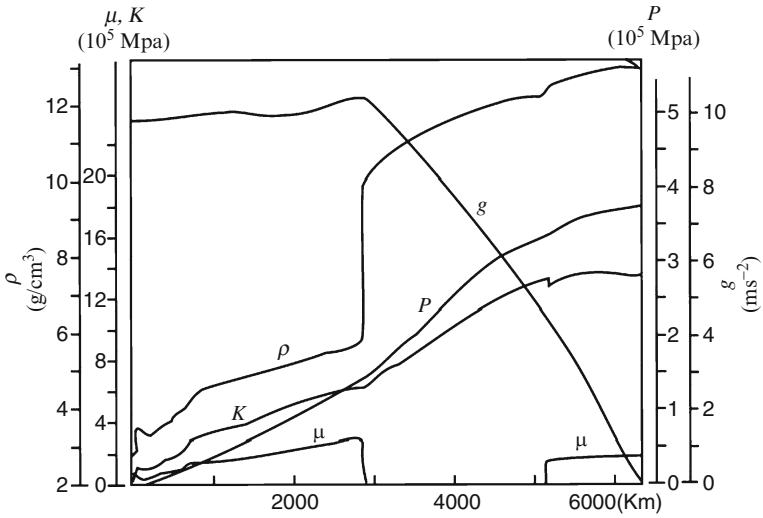

Fig. 2.4 The preliminary Earth reference model. The radial variation of the parameters cited in the text is shown. The surface is indicated by the level 0. Data available from http://solid_earth.ou.edu/prem.html

[2] Pressure (P) waves travel at the greatest velocities within solids and the particle motion is parallel to the direction of wave propagation. Shear (S) waves are transverse waves.

2.1.1.1 Inner Core

The core was the first internal structural element to be identified. Oldham (1906), by studying earthquake records, first suggested that the Earth must have a molten interior. Temperatures at the Earth's centre can reach to 5,000–6,000°C. However, the inner core remains solid due to the extremely high pressure overcoming the effect of high temperatures. It is composed of mainly nickel–iron alloy and some lighter elements (probably sulphur, carbon, oxygen, silicon and potassium). The inner core seems to be rotating faster than the Earth's surface (Glatzmaier and Roberts 1996).

2.1.1.2 Outer Core

The outer core is in liquid state, with convection as the main form of energy transport. The heat necessary to drive convection is derived from the gradual growth of the inner core. As the deeper outer core cools, liquid iron slowly solidifies, releasing heat in the process. The newly formed particles of solid iron also heat the outer core frictionally as they settle down to join the inner core.

The rotating fluid in this layer is responsible for producing the Earth's magnetic field via a dynamo process. Our planet possesses one of the strongest magnetic fields of the Solar System.[3] This is due to the combination of a relatively high rotation rate (24 h) and the thickness of the outer core, where we have fluids in motion. The magnetic field is practically a dipole, with the axis inclined approximately 11.3° from the planetary axis of rotation.

2.1.1.3 Mantle

The mantle lies roughly between 30 and 2,900 km below the Earth's surface and occupies about 70% of the Earth's volume (see Fig. 2.3). Temperatures range between 1,000°C at upper boundary and over 4,000°C at the boundary with the core. Although these temperatures far exceed the melting points of the mantle rocks at the surface, particularly in deeper layers, the materials are almost exclusively solid. The enormous lithostatic pressure exerted on the mantle prevents them from melting, because the melting temperature increases with pressure.

In the upper mantle there is a major area, the asthenosphere, where the temperature and pressure are at just the right balance so that part of the material melts. The rocks become soft plastic and flow like warm tar. This layer is 100–200 km thick and its top is about 100 km below the Earth's surface. It is mainly composed of oxygen, silicon, iron, aluminium and magnesium. The amount of water in the mantle is still a matter of debate (Bolfan-Casanova 2005).

[3] The strength of the field at the Earth's surface ranges from less than 30 μT (0.3 gauss) in an area including most of South America and South Africa to over 60 μT (0.6 gauss) around the magnetic poles in northern Canada and southern Australia, and in parts of Siberia.

Most of the heat generated is transported through the mantle via convection, providing an effective way to transport material and heat from the deep interior.[4] Evidence for the descent of crustal slabs into the mantle indicates that materials travel both ways. The Rayleigh number for the convecting mantle is

$$\mathrm{Ra} = \frac{\mathrm{g}\alpha(\mathrm{T_m} - \mathrm{T_s})(\mathrm{R_m} - \mathrm{R_c})^3}{\kappa\nu},$$

where g is the acceleration due to gravity, α the coefficient of thermal expansion, ν the viscosity, κ the thermal diffusivity, R_m and R_c the outer and the inner radii of the mantle, respectively and T_m and T_s the temperatures of the mantle and surface, respectively. Convection exists for Rayleigh numbers greater than 10^3. For the mantle, values ranges between 10^6 and 10^8.

The mantle heat flow is parametrized in terms of this Rayleigh number Ra.

$$\mathrm{Q_m} = \frac{\mathrm{k}(\mathrm{T_m} - \mathrm{T_s})}{(\mathrm{R_m} - \mathrm{R_c})}\left(\frac{\mathrm{Ra}}{\mathrm{Ra_{cr}}}\right)^{\beta},$$

where k is the thermal conductivity, Ra_{cr} the critical value for the onset of convection and β an empirical constant.

2.1.1.4 Lithosphere

The lithosphere is the outermost shell of the planet. It includes the crust and the uppermost mantle, which are joined across the Mohorovic layer. The solid crust of the Earth floats on top of the upper mantle and has two main components: oceanic and continental, the latter being thicker. Figure 2.5 illustrates the relative abundance of chemical elements in the upper continental crust. Rock-forming elements are the most abundant. Oxygen and silicon compose approximately 72% of the rocks, with the rest being aluminium, iron, calcium, magnesium and sodium.

2.1.1.5 Energy Budget

The internal energy of our planet has two main sources: (1) the potential energy acquired during the accretion process and the energy added by impacts during the Earth's initial growth and (2) the energy released by the radioactive decay of elements such as ^{40}K, ^{238}U and ^{232}Th through the following reactions:

$$\begin{aligned} {}^{238}\mathrm{U} &\longrightarrow {}^{206}\mathrm{Pb} + 8\,{}^{4}\mathrm{He} + 6\mathrm{e}^- + 6\overline{\nu} + 51.7\,\mathrm{Mev} \\ {}^{232}\mathrm{Th} &\longrightarrow {}^{208}\mathrm{Pb} + 6\,{}^{4}\mathrm{He} + 4\mathrm{e}^- + 4\overline{\nu} + 42.8\,\mathrm{Mev} \\ {}^{40}\mathrm{K} + \mathrm{e}^- &\longrightarrow {}^{40}\mathrm{Ar} + \overline{\nu} + 1.513\,\mathrm{Mev}, \end{aligned}$$

where $\overline{\nu}$ are the antineutrinos.

[4] Mantle materials are poor conductors of heat.

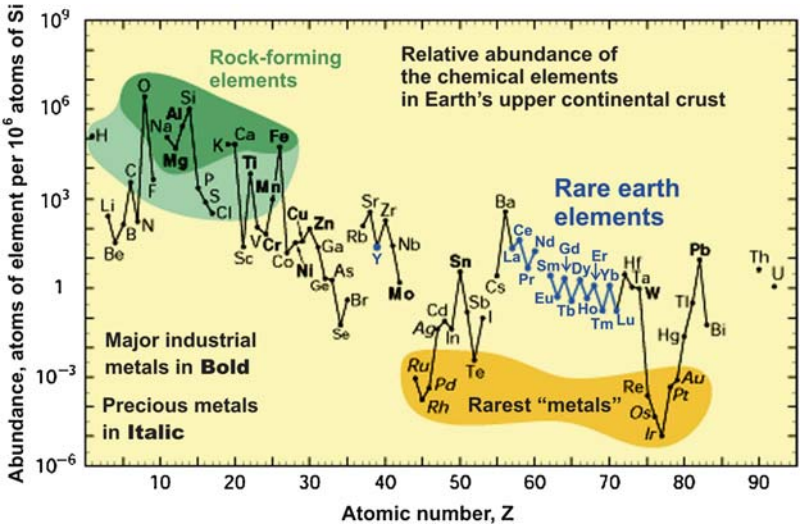

Fig. 2.5 Abundance (atom fraction) of chemical elements in the upper lithosphere as a function of the atomic number

The total power dissipated by the Earth's interior is estimated to be between 30 and 44 TW (Pollack et al. 1993; Garzón and Garzón 2001).[5]

Both types of energy, accretion and radiogenic, decline with time and are proportional to the mass of the planet.[6] They are dissipated, gradually and abruptly, toward the outer layers, giving rise to processes that have configured the structure of the Earth's crust (Condie 2004b). These values are small compared with the 174 PW ($\sim$340 W m^{-2}) received from the Sun at the top of the atmosphere.

The fundamental energy equation for the transfer of energy per unit volume is

$$C\frac{\partial T}{\partial t} = Q_c + Q_H - Q_{conv},$$

where C is the heat capacity of the Earth and Q_c is the heat transfer by conduction, which can be expressed by the Fourier law

$$Q_c = -\kappa \nabla^2 T,$$

κ being the thermal conductivity. Taking into account only Fourier's conductive cooling, W. Thompson (Lord Kelvin) estimated in 1863 the age of the Earth in 100 Ma (Richter 1986; Burchfield 1990). We clearly need other energy sources and cooling mechanisms.

[5] The heat flow is larger in the oceans than in the continents.

[6] The amount of generated energy is proportional to the planetary volume ($\propto R^3$), and the amount of dissipated energy depends on the planetary surface ($\propto R^2$). Therefore, it takes a certain time to cool.

The mantle is a viscous fluid and the heat advented by mass transfer with velocity V, convective heat flux, across the mantle is given by

$$Q_{conv} = c_p V \nabla T.$$

In 1895, John Perry (1850–1920) produced an age of Earth estimate of 2–3 billion years old using a model of a convective mantle and thin crust. However, the main event was the discovery of radioactivity by Henry Becquerel (1852–1908), in 1896, leading to the possibility of longer ages not only for the Earth but also for the Sun and the totality of the Solar System.

Q_H is the local heat generation by radioactivity decaying with time. It depends on the mass, m, of the main radioactive isotopes (expressed in units of 10^{17} kg.)

$$Q_H(TW) = 9.5m(^{238}U) + 2.7m(^{232}Th) + 3.6m(^{40}K)$$

From theoretical estimates we know that $m(^{232}Th):m(U):m(^{40}K) = 4:1:1$. Therefore,

$$Q_H(TW) = 24m(^{238}U).$$

The Urey ratio, U, is defined as

$$U = \frac{Q_H}{Q_{conv}}.$$

Mantle convection models typically assume U values from 0.4 to almost 1 (implicating a surface heat flux of 46 TW), whereas geochemical models (e.g. Korenaga 2008) predict $0.3 < U < 0.5$, implicating smaller heat fluxes (≈30 TW). Latter models assume that U and Th are mainly in the lithosphere and mantle is in the form of oxides, leading to smaller values for the radioactive heat. Other models assume that potassium is alloyed with iron in the interior (Lee and Jeanloz 2003), providing an important source of radioactive energy to generate the magnetic field (Herndon 1996).

Araki et al. (2005) have measured antineutrinos produced by radioactive beta-decay at the heart of the Earth. The results obtained from these so-called geoneutrinos, 19 TW for radiogenic heat, are consistent with compositional models of the planet (Palme and O'Neill 2003; McDonough 2003), and provide a new way of determining where unstable isotopes are stored inside the planet and in what concentrations.

2.1.2 Plate Tectonics

Observed over time, the Earth shows clear changes in the aspect of its surface, which can be easily interpreted as the consequence of internal energy release from its interior. Figure 2.6 shows a computer generated view of the crust relief.

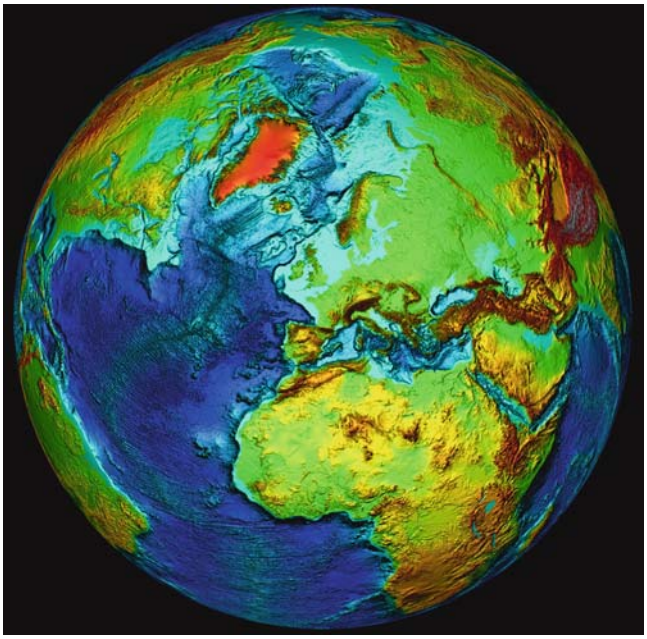

Fig. 2.6 A view of the crust relief showing land and undersea topography. Courtesy of National Geophysical Data Center

During the nineteenth and the early twentieth century, geologists explored the idea that the continents may have moved across the surface.[7] They were inspired by the remarkable fit between Atlantic coasts of Africa and South America, already noted by Francis Bacon.[8] This hypothesis was first developed by Alfred Wegener (1880–1930), who also studied the distribution of animals and fossils to help him in his interpretations.

After World War II, large chains of underwater volcanoes were discovered, known as mid-ocean ridges. The Mid-Atlantic Ridge was mapped first in some detail by M. Ewing (1906–1974) and B. Heezen (1924–1977). These rifts were later identified as newly formed sea floor that is extruded along the ridges. Once emerged, the new floor expands to the sides, process known as sea-floor spreading (Hess 1962).

In principle, there are three primary modes for releasing the heat contained in the Earth's interior, cooling the planet: magma ocean (see Sect. 2.3.1), stagnant lid[9] and

[7] In a letter written on September 1782 to the Abbe Soularie, Benjamin Franklin (1706–1790) recognized that the crust was a shell, which could be broken and parts moved about.

[8] In 1620 he said, *if the fit between South America and Africa is not genetic, surely it is a device of Satan for our confusion*.

[9] In stagnant lid convection, heat is transported by conduction in most of the top layer. The convectively unstable bottom is restricted to this region, which is significantly colder than the interior and not so cold that it is too stiff to participate in the convection.

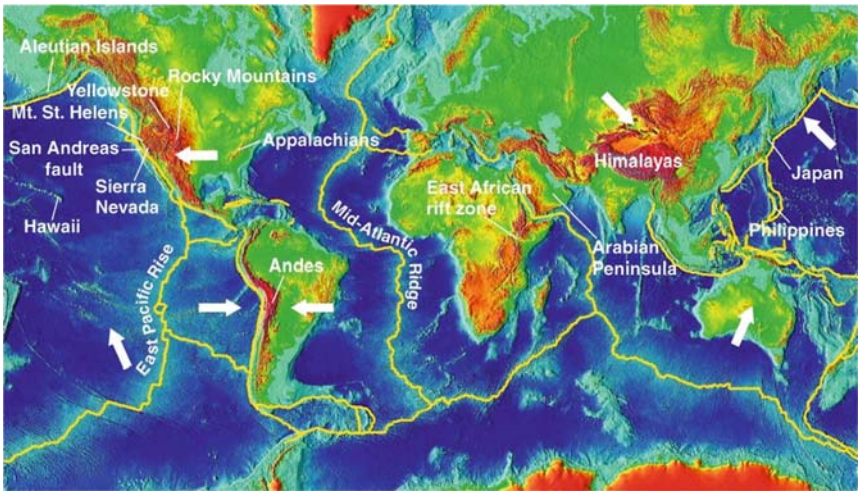

Fig. 2.7 Global plates of the planet

plate tectonics (Schubert et al. 2001). The latter process has been observed only in our planet[10](see Martin et al. 2008 for a didactic summary).

The lithosphere essentially floats on the top of the mantle, the plastic asthenosphere, and is broken up into what are called tectonic plates, which are moving in relation to one another, continuously changing shape and size. The lithosphere is divided into about 20 rigid plates (Fig. 2.7). Dissipation of heat from the mantle is the original source of energy driving plate tectonics. Earth activity (earthquakes and volcanoes) is concentrated at the plate boundaries.

Three types of plate boundaries exist, differentiated by the way the plates move relative to each other. They are also associated with different types of surface phenomena. The different types of plate boundaries are the following:

Divergent boundaries: Here two plates slide apart from each other (examples of which can be seen at mid-ocean ridges and active zones of rifting). Material from the mantle rises from beneath a mid-ocean and partially melts, forming magma and creating new ocean crust.

Convergent boundaries: Two plates move together forming either a subduction zone (if one plate moves underneath the other) or a continental collision (if both plates contain continental crust). Deep marine trenches are typically associated with subduction zones. Because of friction and heating of the subducting slab, volcanism and earthquakes are almost always closely linked to convergent boundaries. The sinking of lithosphere in subduction zones provides most of the force needed to drive the plates and cause mid-ocean ridges to spread.

[10] Venus has a thick lithosphere and mantle plumes (stagnant lid convection) and although plate tectonics may have existed in the past in Mars, it now has become a 'single-plate' planet dominated by hot-spot volcanism (see also Sect. 2.3.2).

Transform boundaries: These occur where plates slide or, perhaps more accurately, grind past each other along transform faults. The relative motion of the two plates is either sinistral (left side toward the observer) or dextral (right side toward the observer).

The destruction (recycling) of crust takes place along convergent boundaries where plates are moving toward each other, and sometimes one plate sinks (is subducted) under another.

2.1.3 The Atmosphere

The atmosphere is the gas layer surrounding the planet (Fig. 2.8). The average pressure at the ground is 1,013 mbar, with a column atmospheric mass of 1 kg cm^{-2} and a density of 2.7×10^9 molecules cm^{-3}. Table 2.2 gives the best estimates about its main constituents.

Oxygen in Earth's atmosphere is very abundant. However, it is relatively rare at the cosmic scale ($\sim$0.06%) and therefore also in the protosolar nebula. Moreover, the noble gases (Kr, Ne) are thousands of times more abundant in the Sun than in our atmosphere. Some mechanism has blown up the primordial atmosphere, leaving the inert nitrogen as the only residual of those early times. We speak in Sect. 2.3.1 on the origin of carbon dioxide and water.

Fig. 2.8 Image of the Earth atmosphere taken on 20 July 2006 from the International Space Station by the astronaut Jeffrey Williams using a digital camera equipped with a 400 mm lens. Astronaut photograph ISS013-E-54329. Courtesy: NASA JSC

Table 2.2 Principal constituents of the atmosphere

Constituent		Composition by volume (%)
Nitrogen	N_2	78.08
Oxygen	O_2	20.95
Argon	Ar	0.93
Carbon dioxide	CO_2	0.0375
Water vapour	H_2O	0.001-4
Neon	Ne	0.0018
Helium	He	0.0005
Methane	CH_4	0.00017
Nitrous oxide	N_2O	0.00003
Hydrogen	H_2	0.00005
Xenon	Xe	0.000009
Ozone	O_3	0.000004

Measurements from balloon soundings and aircraft flights, together with theoretical developments, brought a clearer understanding of the physical structure of the terrestrial atmosphere, which was divided in different layers according to the different processes taking place in them (see Fig. 2.9).

Troposphere: Named after the Greek word for overturning, it extends from the terrestrial surface up to approximately 11 km. The heat source is infrared radiation emitted from the surface, warmed by visible solar radiation. The heat is transferred from the surface to the troposphere by the following processes:

- The evaporation of water and the release of latent heat[11] through the formation of clouds.
- Infrared emission and absorption by greenhouse gases, such as water vapour, CO_2 and CH_4.
- Sensible heat flux, the heat absorbed or transmitted by a substance during a change of temperature that is not accompanied by a change of state.

Assuming hydrostatic equilibrium ($dP(z)/dz = -\rho(z)g(z)$) and the equation of state for an ideal gas ($P = n\,k\,T$), we obtain the following expressions:

$$\frac{dP(z)}{dz} = -\frac{P(z)\mu(z)g(z)}{kT(z)}$$

$$P(z) = P_0 e^{-mgz/kT}; n(z) = n_0 e^{-z/H},$$

where H (z) is the pressure scale height given by

$$H = \frac{kT(z)}{\mu(z)g(z)}$$

[11] Heat absorbed during a change of state.

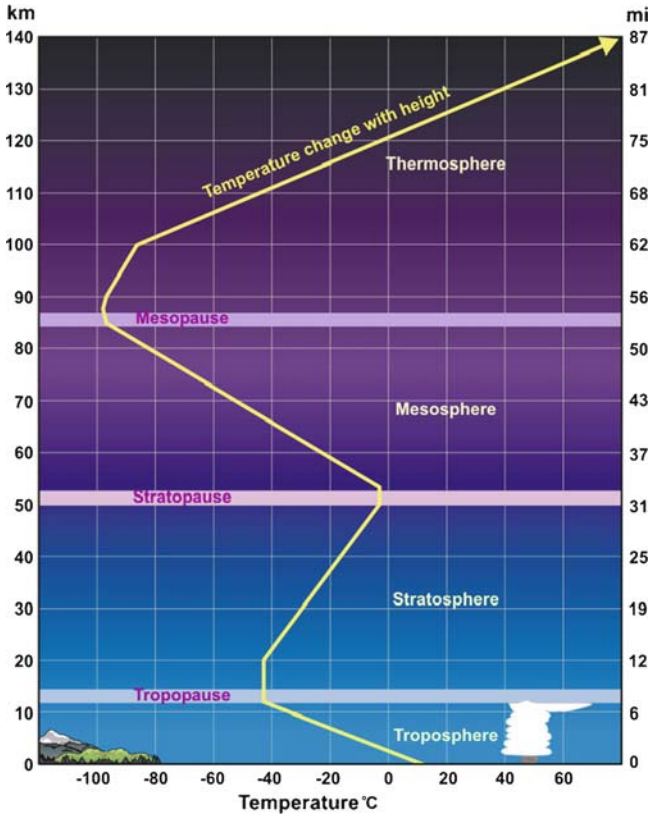

Fig. 2.9 Thermal profile of the atmosphere showing the different layers. Height (in kilometres and miles) is indicated along each side. Credit: National Weather Service

T being the temperature, z the geometrical altitude, p the pressure, g the gravitational acceleration, $\rho = nm$ the mass density, k the Boltzmann constant and μ the mean molecular weight.

Convection is the dominant mechanism for energy transport in the troposphere. Assuming adiabatic motion of the convective cells, the temperature gradient, dT/dz, can be derived by applying the first law of thermodynamics and the consideration of latent heat

$$\mathrm{dT/dz} = -(\mathrm{g(z)/c_p})/[1 + (\mathrm{L/c_p})(\mathrm{dW/dT})] \approx 6.5^{\circ}\mathrm{C/km},$$

where z is the vertical coordinate, g the gravitational acceleration, c_p the specific heat at constant pressure, W the mass of saturated air and L the latent heat of vapourization. Moisture can decrease dT/dz by releasing latent heat. James P. Espy (1785–1860) first derived this parameter empirically for dry and saturated conditions, and some years later verified theoretically by W. Thompson (Lord Kelvin).

The troposphere contains 99% of the water vapour in the atmosphere and its content decreases rapidly with height in this layer. Water vapour plays a major role in regulating air temperature because it absorbs solar energy and thermal radiation from the planet's surface.

The tropopause is highest in the tropics (∼16 km) and lowest in the polar regions (∼8 km), and also undergoes seasonal changes. Here, radiative processes start to dominate. Meteorological processes take place in the lower atmosphere (the tropo- and stratosphere).

Stratosphere: This layer lies between 10 and 50 km altitude and it is mainly a radiation-driven environment. The temperature increases with altitude due to the absorption of UV radiation by ozone, a topic that will be dealt with later (Fig. 2.26). Other major absorbing and emitting gases in this region are carbon dioxide and water vapour (see, e.g. Taylor 2003 for a summary).

Mesosphere: This is the coldest of the atmospheric layers and is created by the emission of radiation from carbon dioxide, CO_2. The temperature decreases with the altitude and reaches low enough values to freeze water vapour producing ice clouds, also called noctilucent clouds. Because of oxidation processes and the penetration of UV radiation, which dissociates polyatomic molecules, this layer is more complex than those below.

Thermosphere: The temperature rises in this layer again because of the heat released from the dissociation of molecular oxygen by UV light and photoionization by X-rays. Here, conduction is the main mode of energy transport. In this layer the absorption of solar energy is less than 1% of that in the stratosphere, but the air is so thin that a small increase in deposited energy can cause a large increase in temperature.

Exosphere: A region where most of the particles have enough kinetic energy to escape from the terrestrial atmosphere. The minimum escape velocity from the Earth, the critical escape velocity, is about 11.3 km s^{-1}.

The outer layers of our planet will be described in more detail in Chap. 4. See Chamberlain (1987) and Houghton (2002) for basic concepts on planetary atmospheres.

2.1.4 Energy Balance of the Atmosphere

2.1.4.1 Albedo

The unit-less quantity albedo (Latin for white) is a measure of the overall reflection coefficient of an object. The geometric albedo, p, is defined as the amount of radiation relative to that from a flat Lambertian surface, which is an ideal reflector at all wavelengths. The bond albedo, a, is the total radiation reflected from an object compared to the total incident radiation. For Earth, the bond albedo (the fraction of incoming sunlight that our planet reflects back to space) is 0.29 while the

geometrical albedo is 0.37 (de Pater and Lissauer 2001). Unless otherwise specified in the following, we refer to the bond albedo (see next chapter for more details).

There are many factors in Earth's climate system that influence how much sunlight our world reflects back to space vs. how much it catches and stores in the form of heat. Potential parameters affecting the albedo are volcanic eruptions, changes in surface vegetation and/or desertification (Betts 2000), variations in snow and ice coverage (Randall et al. 1994), and atmospheric constituents such as aerosols, water vapour and clouds (Cess et al. 1996; Ramanathan et al. 1989; Charlson et al. 1992). Albedo changes will be determined by the total effect of the variations in all these parameters. However, these changing parameters will bring along multiple climate feedbacks, which make assessing the exact implications in the albedo a hard task.

2.1.4.2 The Planet's Mean Temperature

Let us assume a purely radiative balance of the Earth's atmosphere. Climate changes are produced by any perturbation to this balance. Figure 2.10 shows a scheme of the different external factors playing a role in the climate system.

Solar radiation is the primary energy source for the Earth's climate. Its flux at the Earth is determined by

$$F_S = \frac{L}{4\pi d^2},$$

where L is the solar luminosity (energy per time unit) and d the distance Earth–Sun.

The Earth's surface, aerosols in the atmosphere and clouds all reflect some of the incoming solar short-wavelength radiation back to space, preventing that energy from warming the planet. Furthermore, about 13% of the solar radiation incident in the atmosphere is Rayleigh scattered, half of this reaching the Earth's surface as diffuse radiation and the other half being returned to space (Houghton 2002). Short-wavelength radiation, usually defined as having wavelengths between 0.15

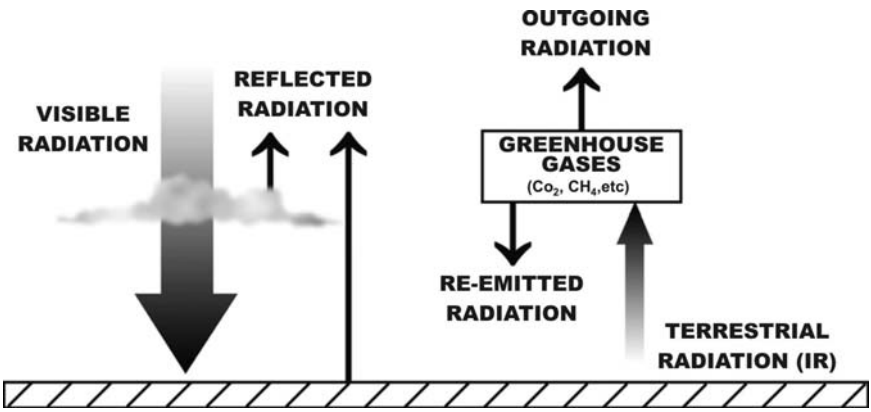

Fig. 2.10 The various factors related to changes in the Earth's surface temperatures

and 4.0 μm, includes about 99% of the sun's radiation; of this energy, 46% is infrared (>0.74 μm), 9% is ultraviolet (<0.4 μm) and the remaining 45% is visible, with wavelengths between 0.4 and 0.74 μm (Liou 2002). A significant portion of the solar energy is absorbed by the Earth (~70%), where it drives terrestrial phenomena before being radiated back into space through the atmospheric window as infrared radiation, peaking at about 10 μm.

The temperature of the Earth is determined by the balance between the received shortwave (visible) flux, F_{in}, and the emitted infrared (long-wave) radiation, F_{out}, so that

$$\left(\pi R_E^2\right) F_S(1-a) = 4\pi\, R_E^2\, F_E.$$

Assuming that the Earth radiates as a black body $F_E = \sigma T^4$, we can define an equilibrium temperature as

$$T_{eq} = \left[\frac{F_S(1-a)}{4\sigma}\right]^{1/4} = \left[\frac{L(1-a)}{4\pi\sigma d^2}\right]^{1/4},$$

where σ is the Stefan–Boltzmann constant and T_{eq} (~255 K) is the equilibrium temperature of the Earth, a physical averaged long-wave emission temperature at about 5.5 km height in the atmosphere (depending on wavelength and cloud cover, altitudes from 0 to 30 km contribute to this emission).

Simple calculations indicate that T_{eq} is much less than the present surface temperature $T_s = 288$ K. Therefore, one must introduce a greenhouse forcing parameter $G[W/m^2]$ defined as the difference between the emission at the top of the atmosphere and the surface. The forcing G increases with increasing concentration of greenhouse gases (see Harries 1996 for an overview on the global energy balance of the Earth).

After Raval and Ramanathan (1989), we can define the normalized greenhouse effect g, with $g = G/\sigma T_s^4$. Then the outgoing power can be written as

$$F_{out} = 4\pi R_e^2 \sigma (1-g) T_s^4.$$

If the planet is in radiative equilibrium, $F_{in} = F_{out}$, then we have

$$T_s^4 = \frac{F_S}{4\sigma(1-g)}(1-a)$$

This means that the albedo, together with solar irradiance and the greenhouse effect, directly controls the Earth's temperature.

2.1.4.3 Greenhouse Gases

Greenhouse gases are the gases present in the atmosphere, which reduce the loss of heat (infrared long-wave radiation) into space by absorbing it and therefore contribute to global temperatures through the greenhouse effect. Two pioneers must be

mentioned in this context: Joseph Fourier (1768–1830) established the concept of planetary energy balance (Fourier 1824), and J. Tyndall (1820–1893) began to study the capacities of various gases to absorb or transmit the heat emitted by the Earth (infrared radiation). He showed that the main atmospheric gases, nitrogen and oxygen, are almost transparent to radiant heat, while water vapour, carbon dioxide and ozone are such good absorbers that, even in small quantities, these gases absorb heat radiation much more strongly than the rest of the atmosphere (Tyndall 1863).

The major atmospheric constituents (nitrogen and oxygen) are not greenhouse gases. This is because homonuclear diatomic molecules[12] such as N_2 and O_2 neither absorb nor emit infrared radiation, as there is no net change in the dipole moment of these molecules when they vibrate. Molecular vibrations occur at energies that are of the same magnitude as the energy of the photons on infrared light. The most important greenhouse gases are diatomic heteronuclear molecules such as water vapour and carbon dioxide (Fig. 2.11); methane, nitrous oxide and other trace gases contribute as well. A simple relation between the surface temperature, T_s, and the partial pressure of carbon dioxide has been given by Walker et al. (1981).

$$\mathrm{T_s} = 2\mathrm{T_e} + 4.6\left(\mathrm{P_t}/\mathrm{P}^0_{\mathrm{CO2}}\right)^{0.364} - 226.4,$$

where t is time in billions of years and

$$\mathrm{T_e} = 255/(1 + 0.087\mathrm{t})^{0.25}.$$

2.1.4.4 2D Models

In zonally averaged climate models, the rate of solar energy input to a latitude belt is locally balanced by the sum of the energy leaving the latitude belt as infrared radiation to space and the net heat transport to other latitude belts. This may be expressed by the relation (North et al. 1981)

$$-\frac{\mathrm{d}}{\mathrm{dx}}\mathrm{D}(1 - \mathrm{x}^2)\frac{\mathrm{dT(x)}}{\mathrm{dx}} + \mathrm{I(x)} = \mathrm{S(x)}(1 - \mathrm{a(x)}),$$

where D is the meridional heat diffusion coefficient, x the sine of latitude, T(x) is the zonally averaged temperature in a given latitude band, I(x) is the outgoing infrared radiation, S(x) is the annual solar radiation reaching the top of the atmosphere and a(x) is the zonally averaged top-of-atmosphere albedo. Caldeira and Kasting (1992) give polynomial fits for these parameters.

[12] Diatomic molecules are molecules made only of two atoms. If a diatomic molecule consists of two atoms of the same element, then it is said to be homonuclear, otherwise it is said to be heteronuclear.

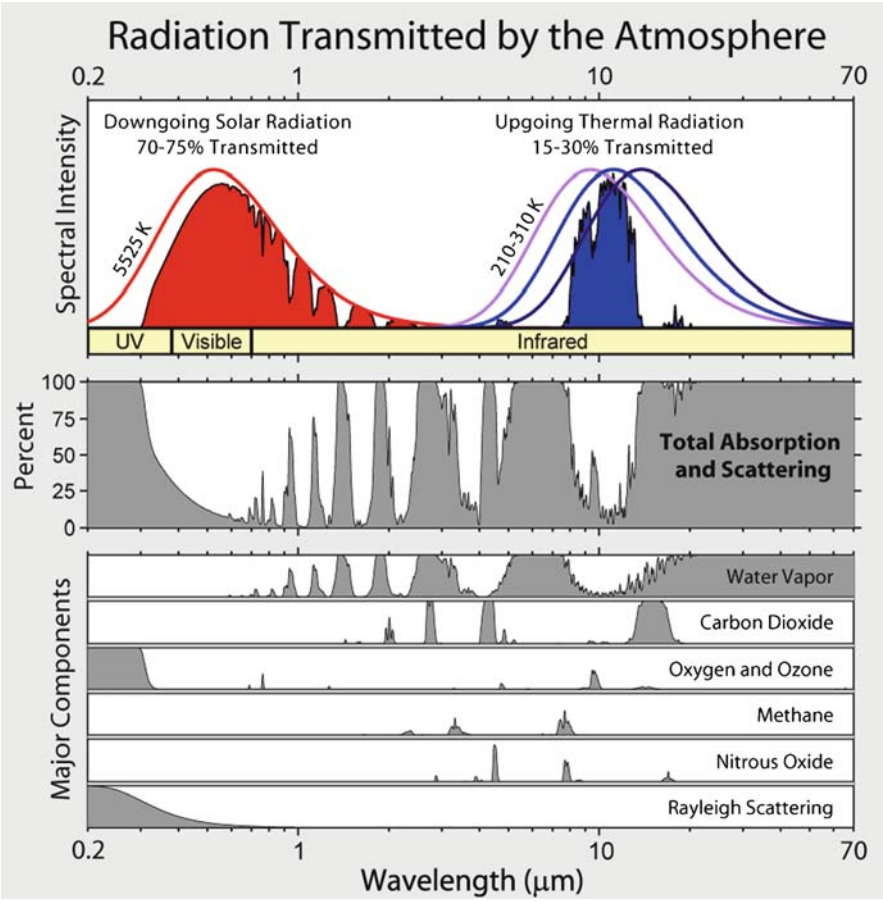

Fig. 2.11 Pattern of absorption bands generated by various greenhouse gases and their impact on both solar radiation and upgoing thermal radiation from the Earth's surface. Note that a greater quantity of upgoing radiation is absorbed, which contributes to the greenhouse effect

2.2 The Precambrian Era (4,500–4,550 Ma BP)

We have now the tools to describe the main epochs of the evolution of our planet. The debate over the age of the Earth and the Sun has been ongoing for over 2,000 years. The discovery of radioactivity near the end of nineteenth century made it possible to clarify the discussions providing a long-standing source for the solar energy and a technique of isotope geochronology. Measurement of the decay of radioactive elements has been applied in meteorites and the Moon, permitting the age of our planet to be estimated at 4,550 millions of years (Patterson 1956; Allegre et al. 1995 and Zhang 2002). The Pb and Hf-W isotopic systems have been widely used for this purpose. It is the beginning of our countdown.

Fig. 2.12 The main events during the Precambrian

HADEAN (4.500 - 3.800)	FORMATION OF CORE MOON FORMATION FAINT YOUNG SUN LATE HEAVY BOMBARDMENT
ARCHAEAN (3800 - 2500)	TECTONIC ACTIVITY ORIGIN OF LIFE (PROCARIOTS) METHANE
PROTEROZOIC (2500 – 550)	RISE ATMOSPHERIC OXYGEN CARBON DIOXIDE CYCLE EUCARIOTS FIRST SNOWBALL EPISODE

Figure 2.12 shows the main periods of the Precambrian era, covering most of the history of the Earth, and the most relevant events occurring during this time period.

2.2.1 The Formation of the Earth: The Hadean Era

Three main phases can be considered in the formation of the Earth (Goldreich et al. 2004). It started with a quick runaway accretion in a disk of small bodies, a process lasting less than one million years. During several hundred million years, the embryos grew at the expense of smaller bodies until finally the orbits of the embryos began to cross, colliding and coalescing in a protoearth (see Chap. 8 for more details).

The Earth's core began to grow after the formation of the protoearth as the temperature increased to the point where dense, liquid iron began to sink toward the centre of the planet. According to a value given by isotopic signatures of radiogenic element pairs $^{182}Hf/^{182}W$ (Kleine et al. 2002), the core formation was completed 30 Ma years after the formation of the Earth. Sometime during this period the surface of the Earth became solid and the first rocks were formed (Wood et al. 2006).

No more than 100 Ma after accretion, the Earth had already reached its present size. Temperatures in the interior were high enough to partially melt the mixed solids of silicate and iron. The differentiation process released a considerable amount of energy. The melting of hot dry mantle at ocean ridges and plumes resulted in a crust about 30 km thick, overlaid in places by extensive volcanic plateaus. The continental crust, in contrast, was relatively thin and mostly submarine.

Figure 2.13 shows a scheme of the evolution of temperature, water and carbon dioxide during the Hadean era. Substantial greenhouse and tidal heating were able to maintain a magma ocean for a few million years.

The Moon would have been formed by a grazing collision with a Mars-mass object during the late stage of Earth accretion (Hartmann and Davis 1975; Stevenson 1987; Canup and Asphaug 2001 and Canup 2004), but with our planet already differentiated into mantle and core (Toboul et al. 2007). The collision formed a dense atmosphere of gaseous silicates that rapidly cooled down and precipitated. The residual atmosphere was constituted by water vapour and carbon dioxide.

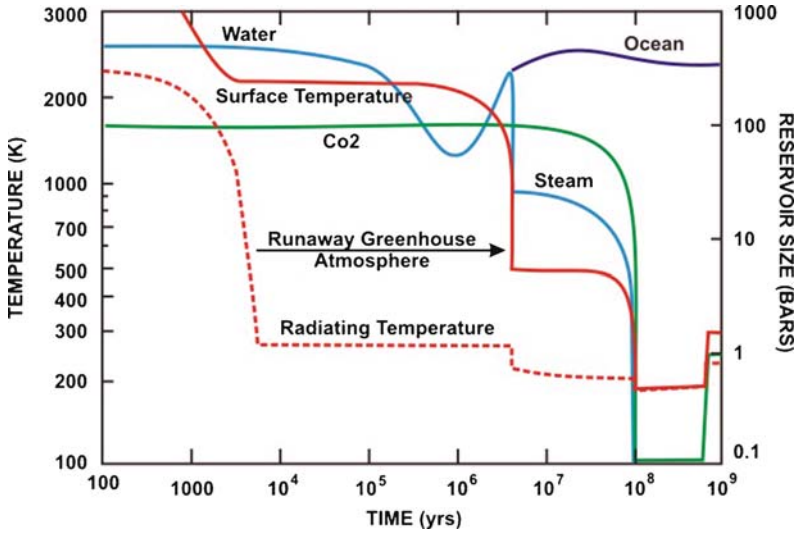

Fig. 2.13 Evolution of temperature, water and carbon dioxide during the Hadean. Source: Zahnle et al. (2007). Copyright: Springer

The partial pressure of CO_2 was between 40 and 200 bars[13] and the temperature around 1,300 K. In a few million years the atmosphere cooled down, the water vapour condensed and precipitated, and at 4.4 Ga BP had already formed a stable ocean (see Wilde et al. 2001 and Pinti 2005).

The building blocks of the Earth were likely dry because temperatures were too high for water to condense or form from hydrated materials. The best alternative source are the bodies of the asteroid belt. Carbonaceous chondrite meteorites originating from C-type asteroids in the outer asteroid belt have Deuterium/Hydrogen ratios similar to our oceans. Models have reproduced the current volatiles inventory via these minor bodies formed in the outer Solar System and gravitationally shifted inward (Morbidelli et al. 2000; Raymond et al. 2004). Comets are not able to fit the D/H ratio of the oceans.[14]

2.2.1.1 The Moon and the Earth Rotation

We will discuss later (Chap. 9) the importance of the Moon for the climate stability of the Earth, and therefore for its habitability. For the moment it is important to understand the processes that originated a satellite so large compared to the mass of the host planet.

[13] Calculated according to the current amount in carbonates, continental sediments and the biosphere.

[14] This parameter has been measured only in three comets: Halley, Hale-Bopp and Hyakutake. The values are factor two larger than the Earth water.

The principle of conservation of angular momentum requires that changes in the Earth's rotation must be produced by the following: (1) torques acting on the solid Earth and (2) changes in the mass distribution within the solid Earth. Here we are interested in the first process produced by the Moon.

The tides are the most important manifestation of the interaction between the Moon and the Earth, producing two ocean bulges, which are instantaneously directly beneath the moon and directly opposite the moon. The Earth's rotation carries the Earth's tidal bulges slightly ahead of the point directly beneath the Moon. This means that the force between the Earth and the Moon is not exactly along the line between their centres, producing a torque on the Earth and an accelerating force on the Moon. This causes a net transfer of rotational energy from the Earth to the Moon, slowing down the Earth's rotation by about 1.5 ms/century and raising the Moon into a higher orbit by about 3.8 cm per year.

Conservation of the angular momentum of the Earth–Moon system implies

$$\mathrm{L} + \mathrm{S} = \mathrm{L}_{\mathrm{tot}} = \text{constant},$$

where S is the spin angular momentum of the Earth and L the orbital angular momentum of the Moon. At present, $L_0/S_0 = 4.83$, but it is important to know the previous history (Arbab 2003).

The Earth's rotational energy is given by

$$\mathrm{E_R} = \frac{1}{2}\mathrm{C}\omega^2$$

and the Moon's orbital energy by

$$\mathrm{E_M} = -\frac{\mathrm{GM_M M_E}}{2\mathrm{r}},$$

where ω, r, $\mathrm{M_M}$ and $\mathrm{M_E}$ are the Earth's angular speed, Earth–Moon distance, Moon mass and Earth mass, respectively. The moment of inertia, C, is given by

$$\mathrm{C} = \frac{\mathrm{L} - \mathrm{L_{tot}}}{\omega}.$$

The rate of total tidal energy, $\mathrm{E} = \mathrm{E_R} + \mathrm{E_M}$, dissipated in the system is

$$\mathrm{P} = -\frac{\mathrm{dE}}{\mathrm{dt}}.$$

The Moon retreats at a rate proportional to its semi-major axis to the $-11/2$ power (as it gets further away, it retreats more slowly). The present rate of lunar recession of 3.82 $\pm$0.07 cm per year was obtained by lunar laser ranging (Dickey et al. 1994). Paleontological evidence comes from tidally laminated sediments. Williams (1990) reported that 650 Ma ago, the lunar rate of retreat was 1.95 $\pm$ 0.29 cm per year, and that over the period 2.5–0.65 Ga ago, the recession rate was smaller (1.27 cm per year). Williams (1997) reanalysed the same data set later, showing a mean recession rate of 2.16 cm per year in the period between now and 650 million years ago.

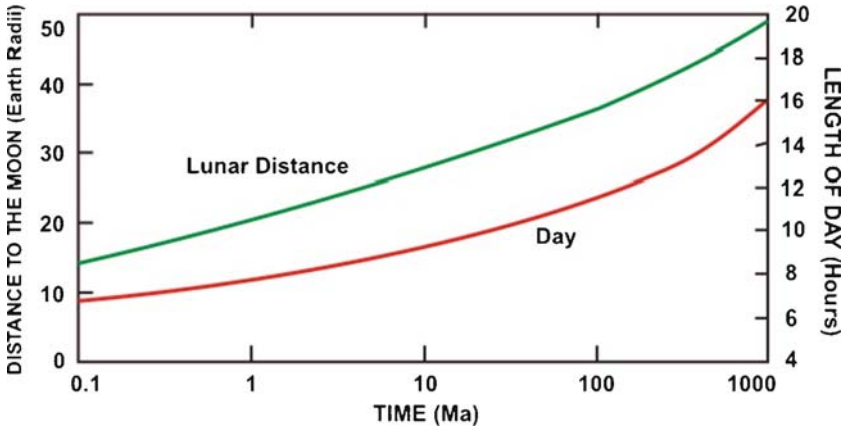

Fig. 2.14 Variation of the distance to the Moon and the length of the Earth's day during the first eon (1 Ga) after the Moon-forming impact. Source: Zahnle et al. (2007) Fig. 9. Copyright: Springer

Sonett et al. (1996) studied sediments of different ages, concluding that the length of the terrestrial day 900 Ma ago was similar to 18 h. Further studies by Williams (2000) in Australian sediments give a mean rate of lunar recession of 1.24 ± 0.71 cm per year during most of the Proterozoic (2,450–2,620 Ma), suggesting that a close approach of the Moon did not occur during an earlier time. Figure 2.14 shows the variation of the distance to the Moon and the length of the Earth's day based on Touma and Wisdom (1998).

The Earth–Moon tidal effects are mutual. The Moon is the smaller object, and the effect of tidal friction has been to change the lunar rotation until its rotational period is equal to its orbital period about the Earth, the present situation. Ultimately, the Moon's action on Earth will produce a similar consequence. When Earth and Moon achieve full synchronism with each rotation being equal to their mutual orbital period, it is estimated that these periods will equal 55 present days and the Earth–Moon distance will be around 610,000 km. For monographs on this topic see Canup et al. (2000) and Lambeck (2005).

2.2.1.2 Late Heavy Bombardment

After planetary accretion, there was a period of quietness in the impact rate on the Earth. Valley et al. (2002) suggested that, during this period (4.4-4.0 Ga), the surface conditions led to liquid oceans and possibly an early emergence of life, which would have been truncated by a catastrophic event. At that time, bodies from the asteroid belt were ejected to the inner solar system by a gravitational perturbation of the outer planets (Strom et al. 2005) (see Chap. 8 for details on the stability of the Solar system).

Radiometric age dating of impact craters on the Moon indicated heavy bombardment with objects larger than 100 km around 3.85 Ga ago, and lasting from

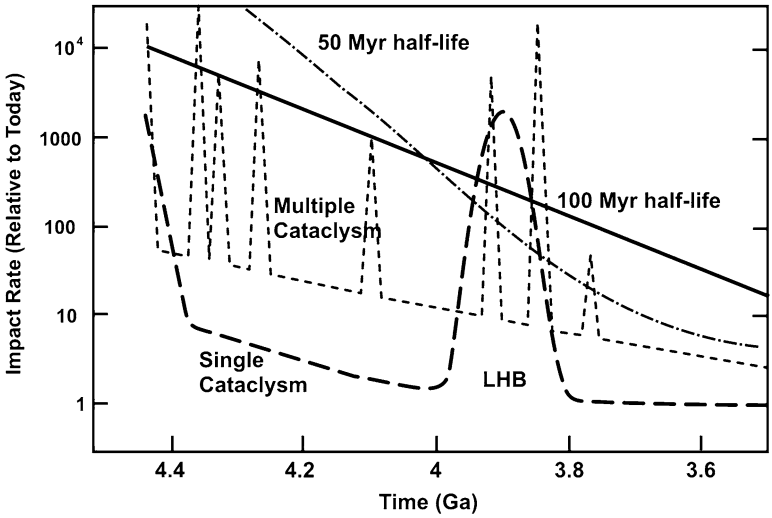

Fig. 2.15 Different concepts of the Late Heavy Bombardment based on Moon data. Continuous process: 50 Ma half-life (Wilhelms 1987); 100 Ma half-life (Neukum et al. 2001); Single cataclysm (Ryder 2002, 2003) and Multiple cataclysm (Tera et al. 1974). Source: Zahnle et al. (2007) Fig. 9. Copyright: Springer

20 to 150 Ma. Computer estimates suggest that, during this period, called the late heavy bombardment (LHB), Earth suffered impacts producing over 22,000 impact craters larger than 20 km, about 40 impact basins larger than 1,000 km, and several continent-sized, 5,000-km basins.[15]

Zahnle et al. (2007) summarizes the different histories proposed for the event. Wilhelms (1987) and Neukum et al. (2001) proposed a declining impact flux that extrapolates back in time. In contrast, Tera et al. (1974) and Ryder (2002, 2003) favoured different types of cataclysms (Fig. 2.15).

The cessation of the LHB coincides well with the first isotopic evidences for life on the Earth ~3.8 Ga ago (Mojzsis et al. 1996). They found that elemental carbon trapped in old rocks of western Greenland have isotopic compositions that span much of the range found in living organisms.

2.2.1.3 The Early Crust and Mantle

Heat flow during the Archaean era was three times larger than today, leading to stronger convection in the mantle and associated higher rates of tectonism. Figure 2.16 shows the variation of the mean heat flow during the first billion years of Earth's history.

[15] Because Earth's effective cross section is 20 times bigger than the Moon, our planet must have suffered many more impacts than those recorded on the lunar surface.

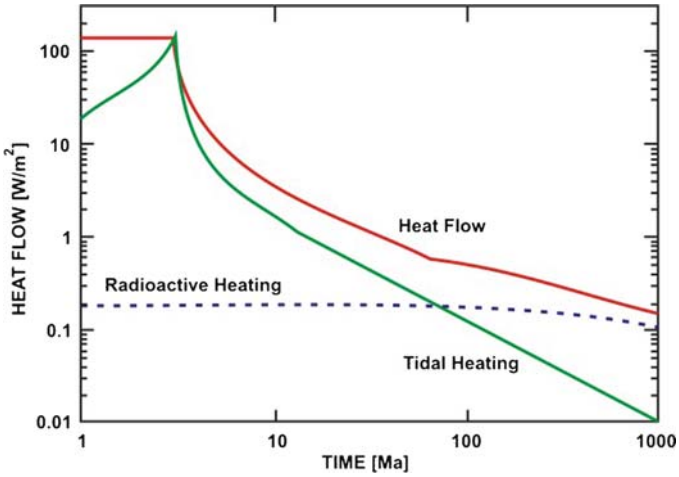

Fig. 2.16 Variation of the mean heat flow from internal energy sources during the first eon of the history of the Earth. Adapted from Zahnle et al. (2007) Fig. 8. Copyright: Springer

A hotter mantle is less viscous, convects faster and releases more heat. This implies that the oceanic crust would be thicker and more buoyant, and therefore difficult to be subducted. For this reason, the appearance of subduction marked the beginning of plate tectonics. There is a strong controversy on the timing, estimates going from the Hadean (Hopkins et al. 2008), late Archaean (Kusky et al. 2001, Sankaran 2006) to the Neoproterozoic (Stern 2005). In any case, the Archaean plates were most likely different from the present ones (De Wit and Hart 1993). An interesting view is that which considers the possibility of intermittency, with transitions from plate tectonics to stagnant-lid tectonics, a less efficient way to remove heat from the mantle. A first consequence would be the existence of jumps in the heat flow (see Sleep 2000; Silver and Behn, 2008).

The oldest rocks on Earth are located in the Isua belt in Greenland (>3.7 Ga), and in the belts of Barberton (South Africa) and Pilbara (Australia), dated in the range of 3.5–3.2 Ga. Zircon grains are the only remnant of the ancient crust.[16]

Cratons are old and stable parts of the continental crust, with deep roots that extend down into the mantle up to a depth of 200 km. The Cratonal lithosphere was formed by processes similar to modern tectonics including subduction (Sleep 2005). The Congo, Kaapvaal, Zimbabwe, Tanzania and West Africa cratons were built between 3.6 and 2.0 Ga ago, and make up for most of the current African continent.

[16] Zircon is a mineral belonging to the group of nesosilicates. Its chemical name is zirconium silicate and its corresponding chemical formula is $ZrSiO_4$.

2.2.1.4 The Young and Faint Sun

The Earth is not an isolated body and all the processes in its atmosphere are strongly dependent on the amount of solar energy received. Therefore, any change in the solar output may have deep consequences in our planet. The variability of the Sun is linked to the dissipation of the available energies (De Jager 1972) . Different sources are characterized by distinct time-scales.

Solar energy is provided by thermonuclear reactions, mostly by the conversion of hydrogen into helium $\left(4^1\mathrm{H} \rightarrow^4 \mathrm{He}\right)$. The mass of the resulting He nucleus is smaller than its constituents and 0.7% of the total mass is converted to energy ($\sim$26.5 MeV)[17].

The thermonuclear fusion process leads to a gradual increase in the molecular weight μ in the core. At present, about 50% of its central H content has been already transformed into He. The thermal pressure is given by $\mathrm{P} \sim \rho \mathrm{R}T/\mu$, where R is the gas constant. Thus, a lower mean molecular weight during the early phases of solar evolution implies a lower temperature (or density) in order to balance the gravitational force. Therefore, the theory of stellar evolution clearly predicts an increase in solar luminosity during its time on the main sequence (Gough 1981). As a consequence, the Sun was 30% dimmer 4.0 Ga ago than at present. This effect can be quantified, as seen below.

Stellar luminosity depends on mass, M, the mean molecular weight, μ, and the radius, R, according to the expression

$$\mathrm{L} \propto \mathrm{M}^{5.5}\mathrm{R}^{-0.5}\mu^{7.5}. \tag{2.1}$$

The process of energy generation in the Sun during its stay on the main sequence is the nuclear transformation of hydrogen into helium. This produces an increase in the mean molecular weight and, according to (2.1), an increase in solar luminosity:

$$\mathrm{L(t)} = [1 + 0.4(1 - \mathrm{t/t_0})]^{-1}.$$

The existence of an early phase of significant mass loss has also been suggested. Wood et al. (2005) obtained high-resolution Lyα spectra and found that the mass loss per unit surface is correlated with the level of magnetic activity. They derived a time dependence of the mass loss of the form

$$\mathrm{dM/dt} = \mathrm{t}^{-2.33\pm0.55},$$

which suggests that the wind of the active young Sun may have been around 1,000 times stronger than that at present. Sackmann and Boothroyd (2003) also proposed the existence of a more massive and larger Early Sun, with the planets being closer with respect to their present positions. In any case, the decay is not necessarily gradual, and a phase of rapid mass loss would lead to the present situation within a period of 1 Ga.

[17] The Sun transforms 4 million tons per second into energy and has lost about 1% of its original mass during its 4.5 billion years of evolution.

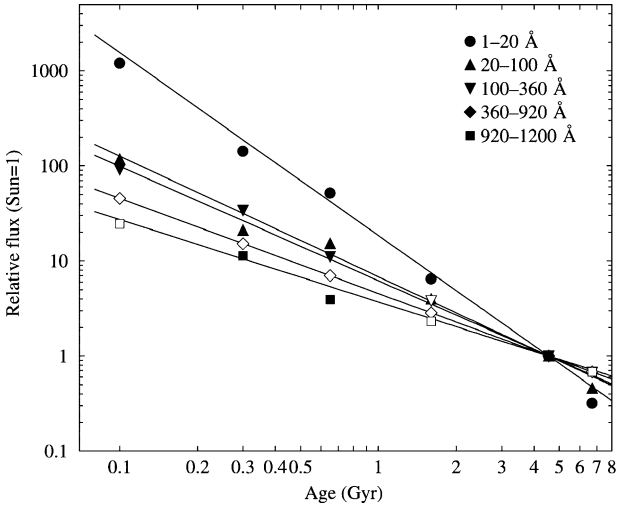

Fig. 2.17 Solar-normalized fluxes vs. age for different stages of the evolution of solar-type stars. Plotted here are the measurements for different wavelength intervals (*filled symbols*) and the corresponding fits using power-law relationships. Represented with empty symbols are the inferred fluxes for those intervals with no available observations. Adapted from Ribas et al. (2005). Reproduced by permission of the American Astronomical Society

The early Sun was not only fainter with respect to present times. Observations of younger Sun-like stars indicate clearly that our Sun was also rapidly rotating. This latter fact implies that the young Sun was a stronger emitter of high-energy radiation (Skumanich 1972; Messina and Guinan 2002). On the basis of the measurements of solar-like stars of different ages, Ribas et al. (2005) have reconstructed the high-energy irradiance in different spectral ranges along the Sun's evolution (Fig. 2.17).

We have already indicated that geological evidence indicates that oceans of liquid water have existed on the Earth at least since 4.4 Ga ago (Wilde et al. 2001), approximately coinciding with the end of the late bombardment period. Moreover, temperatures at the early times were probably much higher than that in our time.

Physical conditions of the atmosphere have changed along the history of the planet and have most likely been an important factor in the mass extinctions. At the geological scale, the climate is controlled by the balance between the variations in the solar output and the changes in the abundance of greenhouse gases in the atmosphere (Fig. 2.18).

Climate modelling indicates that the mean temperature of the Early Earth must have been below zero, a status called 'snowball Earth', from which it would have been impossible to escape (Newman and Rood 1977). This apparent contradiction, called *The faint Sun paradox*, has been explained classically in terms of an enhanced greenhouse effect produced by larger abundances of CO_2 in the primitive terrestrial atmosphere (Dauphas et al. 2007). However, the CO_2 levels were not likely to have been high enough to be the only factor involved in keeping the oceans from freezing (Kasting and Toon 1989; Rye et al. 1995).

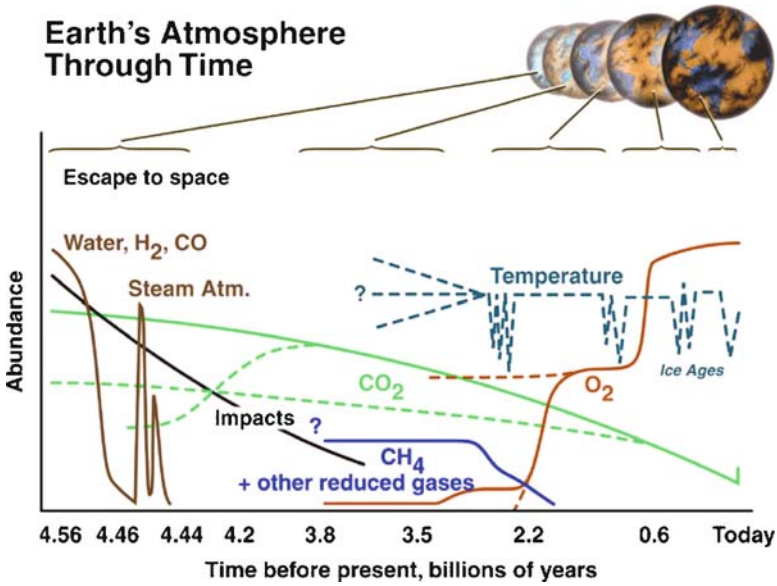

Fig. 2.18 Evolution of the chemical composition of the atmosphere. Adapted from an original drawing of D.D. Marais and K.J. Zahnle. Source: NASA Science News 2002

Güdel (2007) provides an excellent review on this topic. In Chap. 4 we study in more detail the consequences of the magnetic activity of the young Sun on the early atmosphere of our planet.

2.2.2 The Archaean and Proterozoic Times

Two main processes can be associated with the history of our planet during this period. One is related with the geological activity and the recycling of some relevant greenhouse gases, and the second with the effects produced by the appearance of life on its diverse forms. Let us start with the latter event, essential for describing our planet.

2.2.2.1 The Origin and Development of Life

It was first proposed in 1929 by A.I. Oparin (1894–1980) and J.B.S. Haldane (1892–1964) that the synthesis of organic compounds of biochemical significance took place in the primitive terrestrial environment. The prebiotic source of material was formed by contributions from endogenous synthesis in a reducing atmosphere (Miller 1953; Johnson et al. 2008) or neutral atmosphere (Cleaves et al. 2008), metal

sulfide-mediated synthesis in deep-sea vents (Russell et al. 1988) and exogenous sources such as comets, meteorites and interplanetary dust (Oró et al. 2006). For further information about the processes leading to the origin of life, see Schopf (1983, 2002), Delaye and Lazcano (2005) and Chap. 5.

The main features of life's origins and development can be summarized in the following facts (see Schulze-Markuch and Irwin 2004; Ward 2005):

- Life arose relatively quickly
- Life tends to stay small and simple
- Evolution is accelerated by environmental changes
- Once life evolved on Earth, it proved to be extraordinarily resilient.
- Complexity inevitably increases but as the exception not the rule.
- Living things are placed in groups on the basis of similarities and differences at distinct levels.

The cell is the basic structure of life. According to this criterion, life forms were divided by Woese et al. (1990) into three main groups: Bacteria, Archaea and Eucaryotes (Fig. 2.19). All these life forms have a common ancestor, share the same genetic code and biochemistry and developed through the process of evolution. Archaea and bacteria differ mainly in aspects related with the biochemistry and the external parts of the cell. Many archaea are extremophiles. They can survive and thrive even at relatively high temperatures, often above 100°C, as found in geysers and black smokers. Others are found in very cold habitats or in highly saline, acidic or alkaline water. However, other archaea are mesophiles and have been found in environments like marshland, sewage, sea water and soil. Many methanogenic archaea are found in the digestive tracts of animals such as ruminants, termites and humans. Archaea are usually harmless to other organisms and none are known to cause disease. Archaea are usually placed into three groups based on their preferred habitat. These are the halophiles, methanogens and thermophiles.

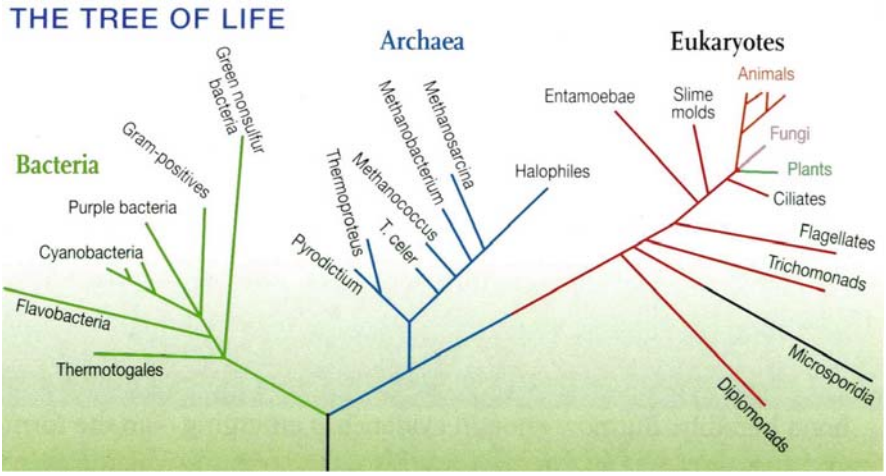

Fig. 2.19 A phylogenetic tree of terrestrial life in three domains proposed by Carl Woese in 1990

Unicellular life is composed of Bacteria and Archaea. About 3.5 Ga ago, Earth temperatures were probably in the range of 55–85°C (Knauth and Lowe 2003). This is the interval where the heat-loving organisms proliferate.

Eukaryotes are organisms in which the genetic material is organized into a membrane-bound nucleus. They appeared in the period 2.1–1.6 Ga ago (Knoll 1992) and were able to sexually reproduce. L. Margulis discovered the mechanism of endosymbiosis within eucaryote cells. Bacteria transformed in organelles[18] within the cell walls.

Unicellular life (microbes) has been the only life form around for close to seven eighths of Earth's history. They still constitute most of the global living mass on Earth today, exerting exclusive control over biomass turnover and biological activity in some Earth regions (deep biosphere, extreme deserts, polar areas etc.).

2.2.2.2 The Carbon Dioxide Cycle

The concentration of greenhouse gases in the atmosphere has been a dominant factor in the evolution of the Earth's climate. Apart from water vapour, carbon dioxide is the most abundant of theses gases in the atmosphere. Therefore, a large part of this chapter will be devoted to the study of its evolution over time.

At geological time scales, the concentration of CO_2 in the atmosphere is controlled by a cycle (Fig. 2.20), first studied by Ebelmen (1845) and Urey (1952). For updated descriptions see Walker et al. (1981), Berner et al. (1983) and Berner (2004). The cycle has two main components: weathering and metamorphism.

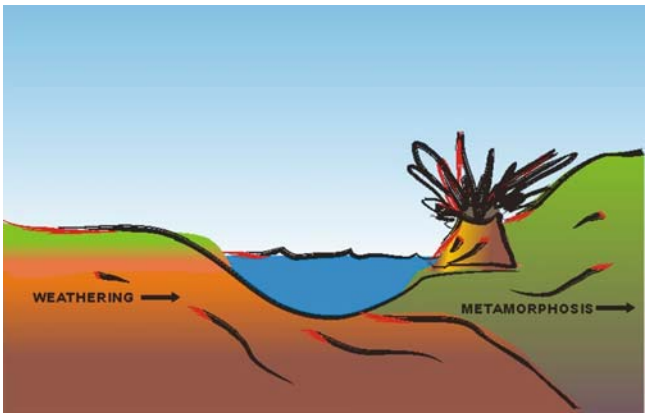

Fig. 2.20 The carbonate-silicate cycle. Adapted from an original drawing of Kasting (1993)

[18] The main organelles are Mitochondria, producing energy from oxygen and food, and Chloroplasts, converting sunlight into energy. Plastids provided eucaryotes with the ability to generate their own oxygen.

1. *Weathering and sedimentation*: Weathering refers to the transfer of carbon from the atmosphere to the rock record and the subsequent marine carbon sedimentation.
First, microbial decomposition in the soils of the continents leads to a buildup of organic acids and CO_2. The carbonic acid weathers the rocks on the Earth's surface, releasing ions of calcium (Ca^{++}) and bicarbonate (HCO_3^-). The reaction is

$$CO_2\,(\text{gas}) + 3H_2O + CaSiO_3 \longrightarrow Ca^{++} + 2HCO_3^- + H_4SiO_4.$$

These products are carried out by groundwater to the rivers and finally to the sea. In the oceans they are precipitated, mostly biogenically, as calcium carbonate

$$Ca^{++} + 2HCO_3^- + H_4SiO_4 \longrightarrow CaCO_3 + CO_2 + H_2O$$

while the silicic acid precipitates as biogenic silica (quartz) following

$$H_4SiO_4 \longrightarrow SiO_2 + H_2O.$$

The whole process can be summarized by the reaction

$$CO_2 + CaSiO_3 \longrightarrow CaCO_3 + SiO_2$$

and similarly for magnesium silicates

$$CO_2 + MgSiO_3 \longrightarrow MgCO_3 + SiO_2.$$

Other subprocesses must also be considered. For example the weathering reaction of calcium carbonate, just the opposite of the precipitation of calcium carbonate into the oceans, which results in no net change of carbon dioxide

$$CaCO_3 + CO_2 + H_2O \longrightarrow Ca^{++} + 2HCO_3^-$$

Berner et al. (1983) adopt the following dependence of the weathering, W, on mean temperature

$$f_W = \frac{W(T)}{W_0} = \left[1 + 0.087(T - T_0) + 1.86 \times 10^{-3}(T - T_0)^2\right],$$

where the subindex 0 indicates present values: $W_0 = 3.3 \times 10^{14}\,\text{g year}^{-1}$ (Holland 1978) and $T_0 = 288\,\text{K}$, the average temperature of the present Earth. An alternative expression is given by Walker et al. (1981) and Caldeira and Kasting (1992)

$$f_W = \frac{W}{W_0} = \left[\frac{a_{H+}}{a_{H+,0}}\right]^{0.5} \exp\left(\frac{T_S - T_{S,0}}{13.7}\right),$$

where a_H+ is the activity of the ion H+, which depends on the concentration of CO_2 in the soil and the temperature.

2. *Metamorphosis*: The calcium carbonate, $CaCO_3$, is eventually subducted down into the Earth (via plate tectonics), where high temperatures and pressures convert it back to carbon dioxide. The process is accomplished by metamorphic decarbonation reactions and by melting with the subsequent release of CO_2 by volcanic activity:

$$CaCO_3 + SiO_2 \longrightarrow CO_2 + CaSiO_3.$$

In the words of Stern (2002): *One can speculate that if subduction zones did not exist to produce continental crust, the large exposed surfaces of rock known as continents would not exist, the Earth's solid surface would be flooded, and terrestrial life, including humans, would not have evolved.*

The carbon cycle has been studied mainly in relation to its role as a thermostat of the global climate (Walker et al. 1981) and the habitability of planetary systems (Franck et al. 2000, 2002). The different phases are described in more detail below.

2.2.2.3 Sea-Floor Spreading and Continental Growth

The dynamic equilibrium of CO_2 fluxes is controlled by the balance between the spreading of the ocean floors, S, the area occupied by the continents, A_c, and the weathering process W (Kasting 1984)

$$\frac{W}{W_0}\frac{A_C}{A_{C,0}} = \frac{S}{S_0},$$

where S is given by

$$S(t) = \frac{q_m(t)^2 \pi \kappa A_{oc}(t)}{[2k(T_m(t) - T_{S,0}]^2}$$

q_m being the heat flux from the mantle, κ the thermal conductivity, k the thermal diffusivity, A_{oc} the area occupied by the ocean floors and T_m and T_s the temperatures of the mantle and the surface, respectively. Obviously at any time the planetary area is $A_E = A_{oc} + A_C$.

Franck et al. (1998) have modelled the evolution of q_m. Its value at the present time is 70 mW m^{-2}, assuming that 20% of the observed surface flow (87 mW m^{-2}) is related to radiogenic heat produced in the continental crust. In the Archaean era, the values of q_m, T_m and S were larger than today's values (by factors 2–3 and 4–9, respectively). Franck and Bounama (1997) discussed the thermal and outgassing history of the Earth for different theoretical models.

For continental growth, two broad types of models are generally assumed: (1) Constant growth with $A_c \propto t$ and (2) delayed growth models, with $A_c = 0$ for $t \leq t_{cr}$ and $A_c \propto t$ for $t > t_{cr}$. These models are summarized in Fig. 2.21. Note that according to Contie (1990), crustal growth had two major pulses in the Archaean and Proterozoic eras.

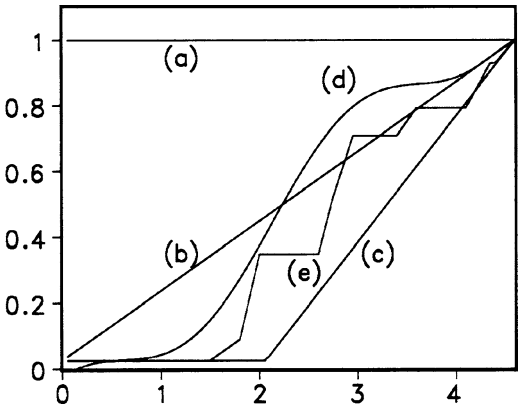

Fig. 2.21 Models of continental growth relative to the present: (**a**) Constant continental surface, (**b**) Linear growth, (**c**) delayed linear growth, (**d**) approximated growth function and (**e**) Episodic growth (Condie 1990). The time is given in Ga starting with at the Earth's origin. Adapted from Fig. 2.5 of Bounama (2007)

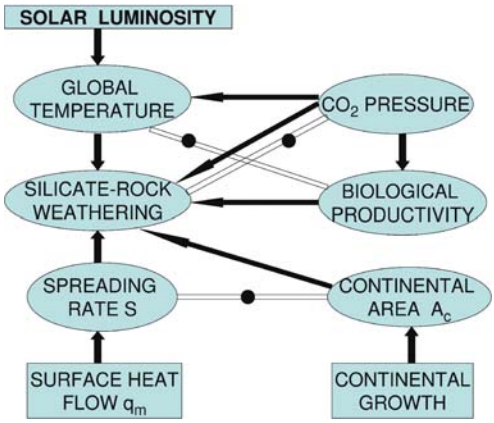

Fig. 2.22 Factors leading to the stability of conditions compatible with life and their mutual influences. *Arrows* indicate direct relationship and arrows with a bullet an inverse one

The balance between the two main processes, outgassing and weathering, has been expressed by Carver and Vardavas (1994) as

$$G(t) = P^{n}_{CO2}(t)A(t)W(t)\exp[(T - 288)/T_c],$$

where T_c is the current temperature (15°C) and n a fitting parameter with values between 0 and 1.

All the above mentioned factors influence the Earth's temperature. Figure 2.22 summarizes them with the main, positive and negative, links.

However, life was not a passive spectator of the CO_2 cycle. The presence of life has clearly influenced the weathering phase of the CO_2 cycle and therefore the environment. Schwartzman and Volk (1989) calculated that if today's weathering is 10,100 or 1,000 times the abiotic weathering rate, then an abiotic Earth would be, respectively, approximately 15, 30 or 45°C warmer than today, suggesting that without biota the Earth today would be uninhabitable.

The biological productivity of photosynthetic organisms, Π, is dependent on the temperature and the partial pressure of CO_2 in the atmosphere. Therefore,

$$\frac{\Pi}{\Pi_{max}} = \Pi(T_S) \,.\, \Pi(P_{CO2})$$

$$\Pi(T) = 1 - \left[\frac{T_S - 50}{50}\right]^2$$

$$\Pi(CO_2) = \frac{P_{CO2} - P_{min}}{P_{1/2} + (P_{CO2} - P_{min})},$$

where $P_{min} = 10^{-5}$ bar = 10 p.p.m is the minimum value of CO_2 pressure required for photosynthesis, and $P_{1/2}$ is the pressure at which Π reaches half the maximum productivity (see Volk 1987).

Therefore, we have

$$f_W(\text{biotic}) = \beta f_W(\text{abiotic}),$$

where

$$\beta = 1 - \sum_{i=1}^{n} \left(1 - \frac{1}{\beta_i}(1 - \Pi_i \, Pi_{i,0})\right),$$

where β is 1 for unicellular organisms and 3.6 for multicellular beings.

The three domains of life reached the maximum productivity at different temperature intervals (Table 2.3). Cooling of the Earth made the emergence of new types possible. Several crucial steps marked this evolution, namely: (1) the colonization of land surface by eucaryotes, (2) Diversification of large, hard-shelled animal life and (3) development of vascular land plants. All of them were associated with important changes in the environment.

Therefore, life probably evolved conditioned by the ambient climate (Schwartzman 1995; Schwartzman and Middendorf 2000). In other words, life evolved opportunistically on Earth in a simple interactive relationship with its environment, sustained by an external energy source, the Sun. The biosphere has been driven forward to greater complexity by the Earth's internal energy resources (plate tectonics). Without these resources and a hydrosphere, it would not be possible to sustain the necessary mass of carbon in the lower crust and mantle to drive the biosphere (Lindsay and Brasier 2002).

Table 2.3 Constants of biological productivity for the three main domains of life. From Franck et al. (2005)

	Procaryotes	Eucaryotes	Multicellular
T_{min} (C)	2	5	0
T_{max} (C)	100	45	30
Π_{max} (Gt/yr)	20	20	20
P_{min} (10^{-6} bar)	10	10	10
β	1	1	3.6

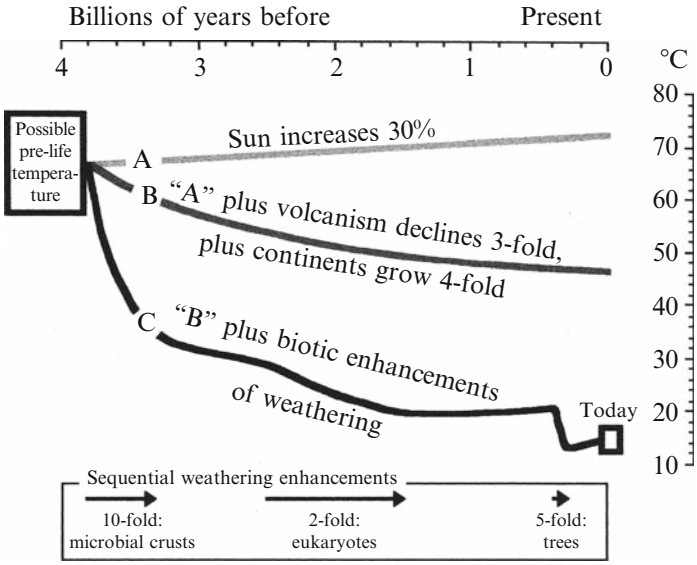

Fig. 2.23 Changes in the global temperature for three different scenarios: (**a**) Solar luminosity increase alone, (**b**) Solar and abiotic cooling and (**c**) Solar, abiotic and biotic cooling. Source: Volk (2003) pg. 237. Copyright: Springer Verlag, New York

Life as a whole has produced an enhancement of weathering over the abiotic rate at a factor 100. Figure 2.23 shows clearly the differences in the biotic and abiotic rates of cooling.

An alternative view suggests an active role of life as a factor of climate variations, as proposed by the GAIA theory (Lovelock 1979). In the original formulation, GAIA was defined as a complex entity involving the Earth's biosphere, atmosphere, oceans and soil; the totality constituting a feedback or cybernetic system, which seeks an optimal physical and chemical environment for life on this planet. The idea has been progressively reformulated and Lovelock (1995) insists more on the co-evolution concepts. Everyone may agree that biota influence the abiotic environment, and that the environment in turn influences the biota by Darwinian processes. Volk (2003) suggested that, once life evolves, Gaia is almost inevitably produced as a result of an evolution towards far-from-equilibrium homeostatic states that maximize entropy production.

A view that is perhaps more difficult to accept states that biota manipulate their physical environment for the purpose of creating biologically favourable, or even optimal, conditions for themselves. After Lovelock and Margulis (1974): *The Earth's atmosphere is more than merely anomalous; it appears to be a contrivance specifically constituted for a set of purposes*.

We agree with Knoll (2003) that the Earth is a biological planet, but not a planet formed by the biota and for the biota. Rather, we live on a planet that was habitable when life first emerged and which has remained habitable through the complex interactions of biological and other components of the Earth system.

The present state of the ecosphere is at the optimum in the sense that the present state has resulted in the emergence of higher life forms, including intelligence. However, this situation will not hold for very long. In a not very distant future (∼1 Ga from now), the increase of solar luminosity will render our planet uninhabitable, at least for multicellular life forms (see Sect. 2.9.1).

2.2.2.4 Greenhouse Gases and Paleoclimate

At very long time scales, the climate of the Earth has been dominated by the increase in solar luminosity and the changes in the abundance of greenhouse gases. Different feedbacks have kept our planet within the limits of habitability. It is a characteristic of stable systems that processes significant on long time scales tend to create negative feedbacks. In the section on the Hadean era, we discussed the problem with the faint young Sun and a possible solution to the paradox through an enhanced concentration of greenhouse gases.

Microorganisms have clearly influenced the evolution of the atmosphere, and this was probably true since early times through the production of a significant amount of methane, another greenhouse gas, contributing to the necessary warming to keep the oceans liquid (Pavlov et al. 2000; Kasting and Siefert 2002 and Kasting and Catling 2003). In fact, methanogenic bacteria are very ancient and the redox state of gases emanating from ancient volcanoes and crust was surely at that time more reduced (Holland 2002).

Methane is destroyed both by photolysis and by reaction with the hydroxyl radical, OH. Hence, in order for it to have been abundant in the early atmosphere, it must have been resupplied by either biotic or abiotic sources (Kasting and Ono 2006). Substrates for methanogenesis should have been widely available,[19] for example

$$CO_2 + 4H_2 \longrightarrow CH_4 + 2H_2O.$$

The photochemical lifetime of methane is relatively short (∼10 years) today, but it would be ∼1,000 times longer in a low oxygen atmosphere. The warmer the world became, the more methane would have been produced. However, the atmospheric abundance of this gas must remain within certain limits. Methane molecules combine together to form complex hydrocarbons, which then condense into aerosols. For this process the resulting climatic effect should be a cooling, lowering the methanogens growth (Pavlov et al. 2001). Figure 2.24 shows the chain of feedbacks keeping the Earth warm at early times.

At that time the Earth had a pinkish-orange colour, similar to that of Titan, the largest moon of Saturn. These hazy skies could have provided a substantial source of organic material, 100 million tons, which would have been useful for the emergence of life on the planet (Trainer 2004).

[19] On the Early Earth, molecular hydrogen, H_2, reached concentrations of up to thousands of parts per million.

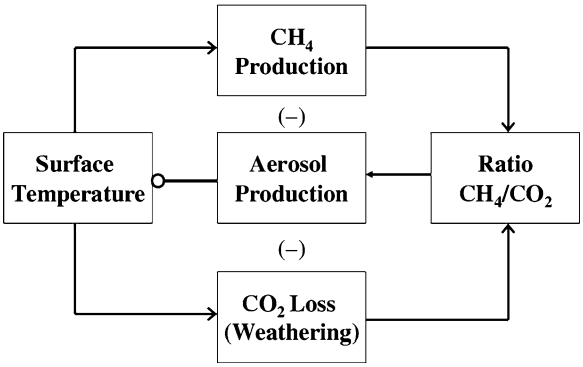

Fig. 2.24 Links between temperature and the abundances of methane and carbon dioxide. Adapted from J. Kasting

The loss of mass-independent fractionation[20] in sulfur isotopes indicate a collapse of atmospheric methane, disappearing abruptly ca. 2.45 Ga ago (Zahnle et al. 2006). The methanogens and their role in the Earth's evolution were soon substituted by other organisms, the cyanobacteria, emitting another gas essential for understanding our planet: oxygen.

2.2.2.5 Oxygen, Ozone and Ultraviolet Radiation

In a primordial atmosphere dominated by carbon dioxide and water vapour, free oxygen atoms can be produced by the following reactions:

$$H_2O + \text{radiation}\,(\lambda < 240\text{nm}) \longrightarrow OH + H$$
$$OH + H \longrightarrow O + H_2O$$
$$CO_2 + \text{radiation}\,(\lambda < 230\text{nm}) \longrightarrow CO + O$$

The free oxygen will produce molecular oxygen and ozone through Chapman reactions (see Chap. 4). On the basis of UV measurements of T Tauri stars, Canuto et al. (1982) indicated that the O_2 surface mixing ratio was a factor 10,000–1,000,000 times greater than the standard value of 10^{-15}. Canuto et al. (1983) extended their calculations to other atmospheric components such as OH, H, HCO and formaldehyde (H_2CO).

At the beginning of the Proterozoic era, the cyanobacteria – photosynthetic procaryotes also known as blue-green algae – brought about one of the greatest changes

[20] It refers to any chemical or physical process that acts to separate isotopes, where the amount of separation does not scale in proportion to the difference in the masses of the isotopes.

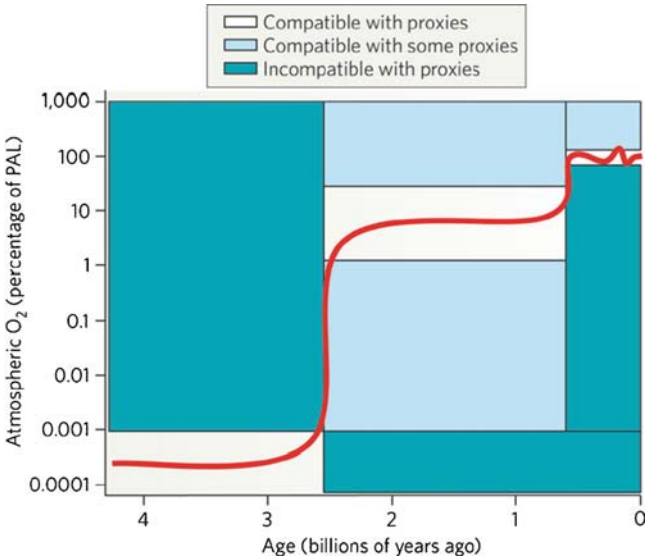

Fig. 2.25 Time evolution of atmospheric oxygen. The *red line* shows the inferred level of atmospheric oxygen bounded by the constraints imposed by the proxy record of atmospheric oxygen variation over Earth's history (Kupp 2008 Fig. 2). Reprinted by permission from Macmillan Publishers Ltd: Nature 451, p. 278, Copyright (2008)

this planet has ever known: a massive increase in the concentration of atmospheric oxygen through the reaction[21]:

$$CO_2 + H_2O \longrightarrow CH_2O + O_2.$$

The evolution of the concentration of oxygen in the atmosphere is not very well known, especially in early times (Kump 2008 and Fig. 2.25). Oxygen photosynthesis arose long before O_2 became abundant in the atmosphere, but oxygen levels in the atmosphere rose appreciably only around 2.5 Ga ago. The signature of mass-independent sulphur-isotope ($^{33}S/^{32}S$) behaviour[22] sets an upper limit for oxygen levels before 2.45 billion years ago (Pavlov and Kasting 2002) and a lower limit after that time. The record of oxidative weathering after 2.45 billion years ago sets a lower limit for oxygen levels at 1% of PAL (Present Atmopsheric Level), whereas an upper limit of 40% of PAL is inferred from the evidence for anoxic oceans during the Proterozoic. The tighter bounds on atmospheric oxygen from 420 million years ago to the present is set by the fairly continuous record of charcoal accumulation (Scott and Glasspool 2006).

[21] Here CH_2O is shorthand for more complex forms of organic matter.

[22] To preserve this signature three conditions are needed: very low atmospheric oxygen, sufficient sulphur gas in the atmosphere and substantial concentrations of reducing gases (methane).

The earliest evidence of anaoxygenic photosynthesis is dated around 3.4 Ga (Tice and Lowe 2004). However, prior to ~2.5 Ga, oxygen did not leave oxidation signatures that are prevalent in the geological record to the present (cf. Canfield 2005), although cyanobacteria were already present since 2.7 Ga (Brocks et al. 2003). This transition is called the 'Great Oxidation Event' by Holland (2002), although it has also received the name of 'Oxygen Catastrophe'.

Kump and Barley (2007) proposed that this transition was favoured by the increase of subaerial volcanoes with respect to the submarine ones, reducing an important oxygen sink. The deep ocean may have become oxic at 1.8 Ga. Instead of a tectonic driver for this transition, Schwartzman et al. (2008) favoured a biological mechanism based on a methane atmosphere producing the emergence of oxygenic cyanobacteria at about 2.8–3.0 Ga. At that time the mean temperature was ~60°C, the most adequate for cyanobacteria.

An apparent paradox arose because, though the terrestrial crust contains 1.1×10^{21} moles[23] of reduced carbon, the total amount of organic carbon to account for all the atmospheric oxygen is only 0.038×10^{21} moles. In other words, the atmosphere contains too little oxygen. Clearly, the missing oxygen is trapped in oxidized reservoirs such as sulphates and ferric iron.

The variations in oxygen content can be explained in terms of a balance between the following factors (Claire et al. 2006):

$$\frac{d}{dt}[O_2] = F_{Sources} - F_{Sinks} = (F_B + F_E) - (F_V + F_M + F_W).$$

$[O_2]$ is the total reservoir of molecular oxygen in units of teramoles, F_B the flux of oxygen due to organic carbon burial, F_E the flux of oxygen to the Earth due to hydrogen escape, F_V and F_M represent flux of reducing gases (i.e. H_2, H_2S, CO, CH_4) from volcanic/hydrothermal and metamorphic/geothermal processes, respectively, and F_W the oxygen sink due to oxidative weathering of continental rocks. Let us illustrate this balance with some examples.

The production of methane inevitably means a greater rate of hydrogen escape, which drives the oxidation of the lithosphere, a lowering of oxygen sinks, the rise of oxygen and, as a consequence, global cooling.

Given a certain amount of oxygen in the atmosphere, the formation of an ozone layer is an unavoidable process through the reactions schematized in Fig. 2.26.

The origin of the ozone layer is probably linked to the Great Oxidation (Goldblatt et al. 2006). The transition was caused by UV shielding, decreasing the rate of methane oxidation once oxygen levels were sufficient to form an ozone layer. As little as 0.01–0.1 present atmospheric levels (PAL, normalized to 1) of molecular oxygen may have been sufficient to produce an effective ozone shield (Kasting and Donahue 1980). The amount of ozone required to shield the Earth from biologically lethal UV radiation, at wavelengths from 200 to 300 nm, is believed to have been in existence ~600 million years ago. At this time, the oxygen level was approximately

[23] One mole of an element is equal to 6.02×10^{23} atoms of that substance.

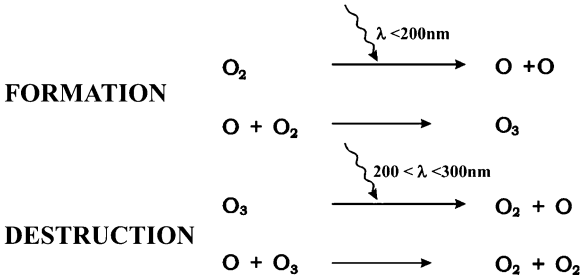

Fig. 2.26 Formation and destruction of ozone, according to the Chapman cycle

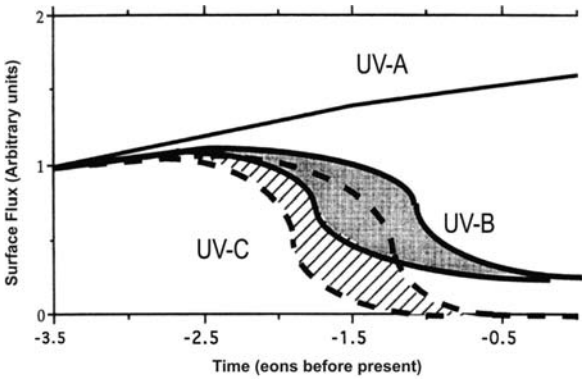

Fig. 2.27 Time variation of the UV radiation reaching the terrestrial surface. The scale is in relative arbitrary units of energy normalized to 1 at 3,500 Ma before present. Courtesy: F. García Pichel, Dept. of Microbiology, Arizona State University

10% of its present atmospheric concentration. Prior to that, life was restricted to the oceans. The presence of ozone enabled multicellular organisms to develop and live on land, playing a significant role in the evolution of life on Earth and allowing life as we presently know it to exist. Rye and Holland (2000) have raised the suggestion that land-dwelling microbes existed at least as early as 2.7 Ga ago. If this is true, our estimate of the abundance of ozone in the stratosphere at that times need to be revised, unless such microorganisms developed a strong defense mechanisms against UV action.

The increased concentration of oxygen has clearly decreased the amount of UV-B and UV-C fluxes arriving at the surface, while the UV-A range has experienced the commented increase of solar luminosity with time.[24] The implications of UV radiation for biological evolution have been studied by Cockell (2000).

García Pichel (1998) considers three main phases in the evolution of UV radiation at the Earth's surface (Fig. 2.27): (1) High environmental fluxes of UV-C

[24] The wavelength ranges of these three spectral bands are UV-C (100–280 nm), UV-B (280–315 nm) and UV-A (315–400 nm).

and UV-B restricting protocyanobacteria to refuges, coinciding with the period of heavy bombardment from interplanetary debris; (2) The appearance of true oxygen cyanobacteria, producing a compound called scytonemin, which screens out the UV radiation but allows through the visible radiation, essential for the photosynthesis; and (3) Gradual oxygenation and the formation of the ozone shield.

A detailed pattern and timing of the rise in complex multicellular life in response to the oxygen increases has not been established. Hedges et al. (2004) suggest that mitochondria and organisms with more than 2–3 cell types appeared soon after the initial increase in oxygen levels at 2,300 Ma. The addition of plastids[25] at 1,500 Ma, allowing eucaryotes to produce oxygen, preceded the major rise in complexity.

2.2.2.6 The Snowball Earth

It has been hypothesized that the Earth was subjected to two main glaciation periods at 2.7 and 0.7 Ga where the ice probably reached the equatorial regions (Harland, 1964). The global mean temperature would have been about –50°C because most of the Sun's radiation would have been reflected back to space by the icy surface. The average equatorial temperature would have been about –20°C, roughly similar to present Antarctica. Without the moderating effect of the oceans, temperature fluctuations associated with the day–night and seasonal cycles would have been greatly enhanced. The glacial deposits at low paleolatitudes have been explained in terms of a global ice cover, 'Snowball' (Hoffman et al. 1998), or ice-free tropical oceans, 'Slushball' (Hyde et al. 2000).

The existence of these events was mainly supported by the occurrence of negative carbon isotopic excursions in glacial marine deposits with thick carbonate (limestone) (see Shields and Veizer 2002 and Fig. 2.28). The carbon isotope ratio, δ^{13}C, is expressed by the following ratio

$$\delta^{13}\mathrm{C} = \left[\frac{^{13}\mathrm{C}/^{12}\mathrm{C}(\mathrm{sample})}{^{13}\mathrm{C}/^{12}\mathrm{C}(\mathrm{standard})} - 1\right] \times 1000.$$

Life used the lighter ^{12}C isotopes more than the abiotic materials. Extinctions should be reflected in larger values of δ^{13}C. The biological productivity of the oceans virtually ceased during the glacial periods. Together with anoxic conditions beneath the ice, this probably annihilated many kinds of eucaryotic life.

The main snowball event[26] was characterized by three broad intervals of widespread glaciation: the Sturtian glaciation, which occurred around 723 Ma, the

[25] Pastids are a group of organelles that play central roles in plant metabolism via photosynthesis, lipid and aminoacid synthesis (Wise and Hoober 2006).

[26] The first event probably took place around 2.5 Ga ago and it is based on glacial deposits in the Gowganda Formation in Canada and the Makganyene Formation in South Africa (Hilburn et al. 2005; Kopp et al. 2005).

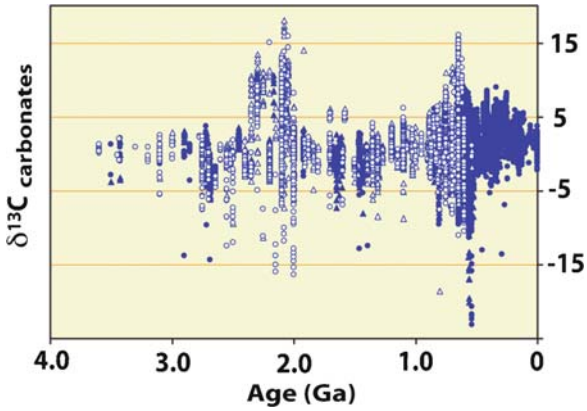

Fig. 2.28 Published δ^{13}C values for marine carbonates (Shields and Veizer 2002). Copyright: American Geophysical Union

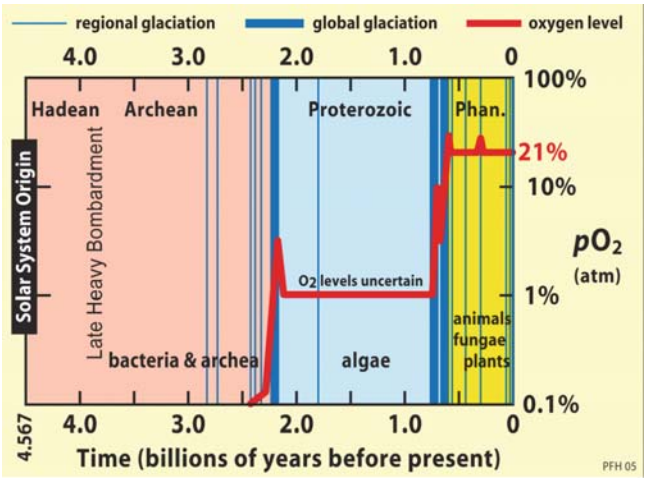

Fig. 2.29 Rise in atmospheric oxygen over geological time and its relationship to the snowball events (*blue bars*) and biological innovations. Courtesy: Paul F. Hoffman

Marinoan (Varanger) glaciation (659–637 Ma) and the Gaskiers glaciation occurring around 680 Ma ago. At that time, large portions of the continental land masses probably were within middle to low latitudes, a situation that has not been encountered at any subsequent time in the Earth's history (Kirschvink 1992). It was also a period of continental dispersal, involving the breakup of the supercontinent Rodinia and the aggregation of megacontinent Gondwana (Hoffman and Schrag, 2002).

The standard interpretation is based on reduced solar luminosity at that time (6% lower) combined with a burst in the oxygen production, leading to a decrease in the concentration of methane and carbon dioxide, two powerful greenhouse gases (Fig. 2.29) (Carver and Vardavas 1994).

Budyko (1969) shows the existence of the different types of stability of the climate system with respect to the ice coverage. When radiative forcing declined, the ice reaches ~30° latitude, after which ice-albedo feedback is self-sustaining and ice lines move rapidly ($<10^3$ years) to the equator (Snowball Earth). After millions of years, dependent on the magnitude of the CO_2 hysteresis loop, normal volcanic outgassing combined with reduced silicate weathering caused CO_2 to reach the critical level for deglaciation. Meltdown occurs rapidly ($<10^4$), driven by reverse ice-albedo and other feedbacks, resulting in an ice-free state with greatly elevated CO_2, taking 10^5–10^7 years for this excess to be eliminated through silicate weathering of the glaciated landscape.

It is a matter of fact that photosynthetic life survived throughout the event, excluding models where thick ice could have prevented sunlight from reaching the underlying ocean. Sedimentary evidence provided by Allen and Etienne (2008) indicates that although ice was probably formed at low-latitudes, some of the Earth's oceans remained ice-free and permitted free exchange with the atmosphere. The planet escaped from a global glaciation probably through the link between the physical system and the carbon cycle. Peltier et al. (2007) suggested that when the atmospheric oxygen at surface temperature was drawn into the ocean, it could have remineralized a vast reservoir of dissolved organic carbon, which in turn caused atmospheric CO_2 levels to increase. As an additional mechanism, Kennedy et al. (2008) suggest that equatorial methane clathrates[27] were destabilized, providing additional warming through methane emission to the atmosphere.[28]

A completely different explanation of the event was proposed by Kirschvink et al. (1997) and Williams et al. (1998), suggesting that continental land masses moved far from the equator. The entire lithosphere reacted to bring them back to the equator much faster than the ordinary tectonic process, with the side effect that the continents have realigned with respect to the magnetic north pole. This hypothesis has been criticized by Torsvik and Rehnström (2001) and Levrard and Laskar (2003).

Stern (2005) proposed that the Neoproterozoic era marked the start of modern plate tectonics, resulting in a major increase in explosive volcanism and a cooling of the Earth's surface. The event was probably accompanied by a polar wander.

2.3 The Phanerozoic Era

The Phanerozoic era covers the Earth's history from 600 Ma to the present. Its start was characterized by a rapid flowering of multicellular life forms, the Cambrian Explosion. At that time the Earth's surface presented the distribution of continental masses shown in Fig. 2.30.

[27] Also called methane hydrate, this is a solid form of water that contains methane within its crystal structure. Its current locations are the continental margins and the permafrost of Siberia and Antarctica.

[28] Allen and Etienne (2008) quoted Louis Agassiz (1807–1873) saying: *the Earth may have avoided death enveloping all nature in a shroud*.

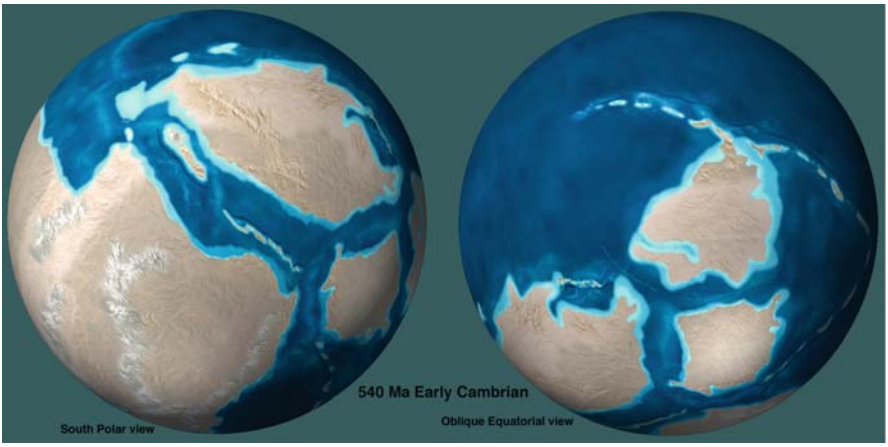

Fig. 2.30 Paleogeographic view of the Earth at the Early Cambrian era (540 Ma ago). Courtesy: Ron Blakey, Department of Geology, Northern Arizona University

The advent of sexual reproduction led to a rapid increase of eucaryotic microalgae between 1,100 and 900 Ma ago. From then to the close of the Precambrian era, microalgal activity declined as atmospheric CO_2 decreased and the climate became colder. Multicellular animals arose during this period, possibly 800–700 Ma ago, and the coelomic kinds that burrow through sediments were efficient producers of organic carbon. Burial of this carbon led to an increase in the oxygen content of the atmosphere (32 g of O_2 for every 12 g of carbon buried), and the subsequent increase of oxygen together with the decrease of carbon dioxide and temperature caused extinction of the microalgae.

The first evidence of complex multicellular animals is given by the so-called Ediacaran biota (McMenamin 1998), some 600 Ma ago, marking the start of the so-called Cambrian explosion, when all the major invertebrate phyla made their appearance. The first macroscopic land plants date back to the Devonian era (400 Ma ago).

Scott et al. (2008) have proposed that molybdenum depletion in the oceans, together with a similar oxygen deficit, may have produced the delay in the development of complex life. This element is used by some bacteria to convert the atmospheric nitrogen in a useful form for living things, a process known as nitrogen fixation.

Watson (2008) supported the hypothesis that complex life is separated from procaryotes by several unlikely steps. Probably this is not in contradiction with other theories that suggest that the Cambrian explosion was triggered by environmental changes (Marshall 2006). This leads us to consider this link in more detail.

The previous mentioned biotic influence of life on the cooling of the planet was amplified in this period by two processes: (1) the diversification of land plants led to increasing chemical weathering of rocks, and therefore an increasing flux of carbon from the atmosphere to rocks, and of nutrients from the continents to the

oceans therefore to thus decreasing CO_2 levels, and (2) the presence of organisms, such as foraminifera, that increased the transportation of carbonates to the oceanic sediments.

2.3.1 The Drift, Breakup and Assembly of the Continents

The movement of plates has caused the formation and break-up of continents over time, including occasionally the formation of supercontinents (see Table 2.4). This process has a cycle of 250–500 Ma. See Rogers and Santosh (2004) and Nield (2007) for monographs on this topic.

In the mid 1960s, J. Tuzo Wilson (1908–1993) showed that continents show a cyclic history of rifting – drifting and collision, followed by rifting again, taking about 500 million years to complete a period. He described the process as a periodic opening and closing of oceanic basins. Plates divert apart and new ocean basins are born, followed by motion reversal, convergence back together and plate collision. Sea level is low when the continents are together and high when they are apart.

The last supercontinent, Pangaea, was composed by two subunits: Laurasia and Gondwana (Fig. 2.31). With the final breakup of Laurasia (60 Ma ago), the continents assumed their familiar configuration of our days.

Table 2.4 List of supercontinents

Name	Period of existence (Ma)
Ur	3,000
Kenorland	2,700
Columbia (Nuna)	1,800–1,500
Rodinia	1,100–750
Pannotia	650–540
Pangaea	300–180
Pangaea Ultima	250–400 Ma from now

The time in brackets is given in Ma

Fig. 2.31 The continents Laurasia-Gondwana at the Triassic, 200 million years ago

Vitousek et al. (1996) modelled the number of mammalian species that would be expected if all the continents were reunited in our days. They concluded that only half of the current species would be present. Plate tectonics clearly promotes biotic diversity.

Silver and Behn (2008) remark that at present most of the subduction zones are in the Pacific basin. At 350 Ma from now, this structure will disappear following the collision of Western America with Eurasia. As a consequence, plate tectonics may temporarily cease. As we have mentioned previously, such intermittency could also have happened in the past. Carbon burial and concomitant oxygen enrichment of the atmosphere could well have been associated with the supercontinent cycle (Lindsay and Brasier 2002).

2.3.2 Supereruptions and Hot Spots

Processes in the Earth's interior may release enough heat to generate plumes of material. As these plumes rise in the uppermost mantle, they spread beneath the lithosphere and begin to melt large plumes of basalt, which erupt in the surface. The mantle plume concept was first proposed by Morgan (1971) and was based on Wilson's (1963) ideas that stationary hot-spots in the shallow mantle underlay island/seamount chains in the deep ocean. Such events may pump large amounts of CO_2 into the atmosphere. According to Condie (1998, 2002), the most important of these events, the superplumes, took place at 2,700, 1,900 and 1,200 million years ago.

Increased production rates of juvenile crust correlate with formation of supercontinents and with superplume events. There may be two types of superplume events: catastrophic events, which are short-lived (<100 Ma), and shielding events, which are long-lived (200 Ma). Catastrophic events may be triggered by slab avalanches in the mantle and may be responsible for episodic crustal growth. Superplume events, caused by the shielding of the mantle from subduction by supercontinents, are responsible for relatively small additions of mafic components to the continents and may lead to supercontinent breakup (Condie 2004).

Inside tectonic plates, we also find transitory emergence of material, forming hot spots. Somewhere between 40 and 150 of these events have been described in both oceanic[29] and continental areas. For a list of hot spots during the last 250 million years, see Rampino and Stothers (1998) and Isley and Abbott (1999).

In summary, two convective processes drive heat exchange within the Earth: plate tectonics, which is driven by the sinking of cold plates of lithosphere back into the mantle asthenosphere, and mantle plumes, which carry heat upward in rising columns of hot material, driven by heat exchange across the core–mantle boundary. The sinking of vast sheets of oceanic lithosphere back into the mantle is the primary driving force of plate tectonics, where the sinking of these slabs is balanced

[29] The Hawaii and Canary Island archipelagos are probably a consequence of two such hotspots.

by the passive upwelling of asthenosphere along mid-oceanic ridges. In contrast, mantle plumes are narrow columns of material that rise more or less independently of plate motions.

2.3.3 The Connection Temperature-Greenhouse Gases

The climate of the late Precambrian was typically cold with glaciation spreading over much of the Earth. At this time the continents were assembled in a short-living supercontinent called Pannotia.

Figure 2.32 show the variation of climate along the Phanerozoic, with a main periodicity of 140 Ma. It oscillates between hot and cold phases, probably reflecting the supercontinent cycle. During the formation of supercontinents we have a lack of sea floor production and a cooler climate. On the other hand, during the break-up process we have a high level of sea floor spreading and high levels of greenhouse gases leading to a warm climate.

More in detail, potential causes for the climate changes at this time scale are the following: (1) Latitudinal position of continents. One key requirement for the development of large ice sheets is the existence of continental land masses at or near the poles. (2) Opening and closing of ocean basins, altering ocean and atmospheric circulation. When there were large amounts of continental crust near the poles, the records show unusually low sea levels during ice ages, because there were many polar land masses upon which snow and ice could accumulate. However, during times when the land masses clustered around the equator, ice ages had a smaller effect on sea level.

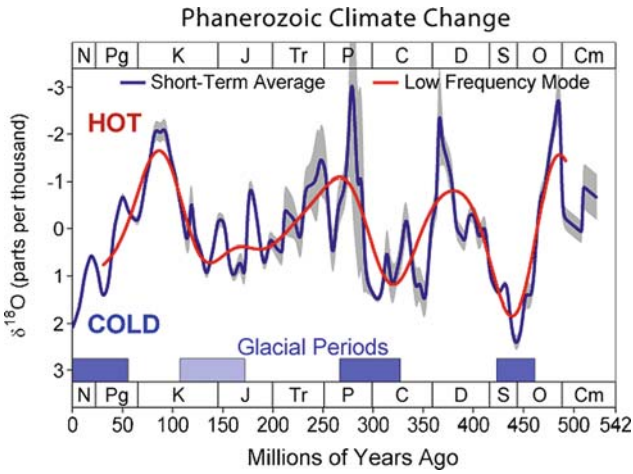

Fig. 2.32 500 million years of climate change based on the data of Veizer (2000). Courtesy: Robert A. Rohde, Global Warming Art

However, the main drivers are the changes in CO_2 content of the atmosphere. Although this has been a matter of debate (Rothman 2002; Veizer et al. 2000; Ryer et al. 2004), recently Came et al. (2007) clearly established that times of minimal CO_2 coincide with the two main glaciations: Permo-Carboniferous 330–270 and the present starting 30 Ma (see also Berner 2004). Also periods of unusual warmth (Mesozoic 250–265 Ma) were accompanied by relatively large concentrations of CO_2 in the atmosphere (see Fig. 2.33). Royer et al. (2007) have estimated that a climate sensitivity greater than 1.5°C to a doubling in CO_2 concentrations has been a robust feature of the Earth's climate system over the past 420 million years. High-resolution CO_2 records of Fletcher et al. (2008) for the Mesozoic and early Cenozoic confirm the leading role of this greenhouse gas in controlling the global climate.

During the Phanerozoic era the oxygen content underwent an important increase during the Carboniferous period reaching values around 35%, accompanied by a decrease in CO_2 with its subsequent biological implications (Graham et al. 1995; Lane 2002).

Figure 2.34 shows the variation of the climate over the last 65 million years. The data are based on oxygen isotope measurements in benthic foraminifera (Zachos et al. 2001). Specially relevant is the Paleocene-Eocene Thermal Maximum, occurring at 55 Ma, a sudden warming of 6°C in only 20,000 years, associated with an increase in CO_2 atmospheric concentration and rise in sea level. The process was probably enhanced by the degassing of methane clathrates, which accentuated a pre-existing warming trend (Katz et al. 2001) . The increasing accumulation of or-

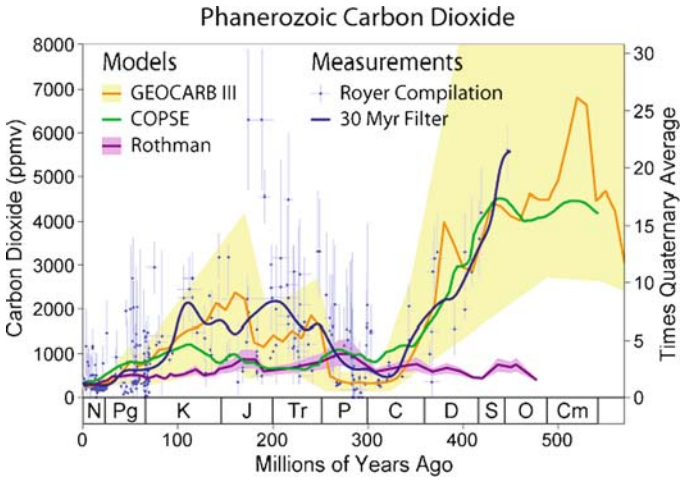

Fig. 2.33 Changes in carbon dioxide concentrations during the Phanerozoic era. Three estimates are based on geochemical modelling: GEOCARB III (Berner and Kothavala 2001), COPSE (Bergmann et al. 2004) and Rothman (2001). These are compared to the carbon dioxide measurement database of Royer et al. (2004) and a 30 Ma filtered average of those data. Error envelopes are shown when they were available. The right hand scale shows the ratio of these measurements to the estimated average for the last several million years (the Quaternary). Courtesy: Robert A. Rohde, Global Warming Art

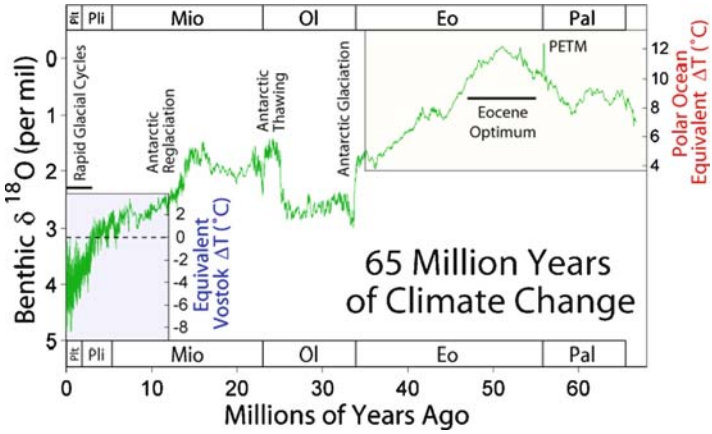

Fig. 2.34 The last 65 million years. Courtesy: Robert A. Rohde, Global Warming Art

ganic material in the deep sea may have cooled this hot climate, starting the current large-scale cooling period (Bains et al. 2000). This cooling is clearly associated with a large decrease in the CO_2 concentration, probably accelerated by the continuous buildup of the Tibetan plateau.

Some geological events during this last phase favoured the evolution towards the emergence of Homo Sapiens through their influence on climate. About 25 Ma ago, continental movements led to the collision of India against the Asian continent, giving rise to the formation of the Himalaya mountains. This originated dry winds blowing to Africa producing a dry season there. It is also worth mentioning that highly encephalized whales, dolphins and porpoises occurred with the drop of ocean temperatures 25–30 Ma ago (Schwartzman and Middendorf 2000).

2.3.4 Temporal Variations of the Magnetic Field

In molten igneous rocks, magnetic particles will align themselves with the Earth's magnetic field. Application of this technique, called paleomagnetism, allows variations in the field to be seen throughout history. The Earth's magnetic field reverses at intervals, ranging from tens of thousands to many millions of years, with an average interval of approximately 250,000 years. The last such event, called the Brunhes-Matuyama reversal, is theorized to have occurred some 780,000 years ago.

The frequency of reversals, the duration over which the reversals occur and the strength of the magnetic field, are well correlated (Glatzmaier et al. 1999). Over long time scales, a quiet magnetic period started 120 Ma ago, lasting 35 Ma. From when it ended (~65 Ma) until the present time, the reversals have become more frequent. Figure 2.35 compares the reversal frequency with massive extinction events (Courtillot and Olson 2007). This correlation is explained in more detail in the following subsection.

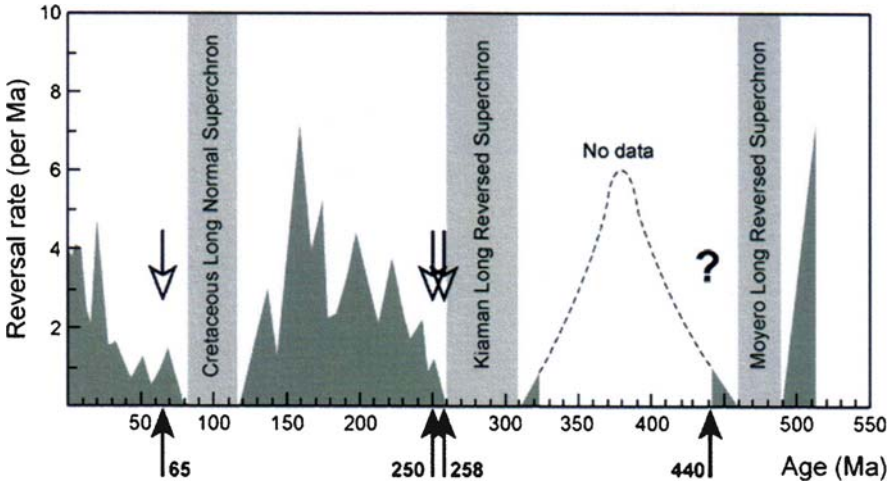

Fig. 2.35 Variation of geomagnetic polarity reversal frequency during Phanerozoic time compared with major faunal mass depletion events (*up arrows*) and associated trap eruptions (*down arrows*). Courtesy: V. Courtillot. Reprinted with permission from Elsevier

2.3.5 Mass Extinctions in the Fossil Record

2.3.5.1 Historical Introduction

The order that characterizes all the structures (including the living beings) existing in the Universe is subject to the conditions of an open thermodynamical system. This is also the case of our Earth. There is a continuous exchange with the media surrounding it, in the form of radiation in all the energy ranges (see Chap. 7 of Vázquez and Hanslmeier 2005) and bodies of different sizes (see Hanslmeier 2008). In this section we study how this interaction has affected, in a direct or indirect way, the evolution of life on our planet (Fig. 2.36). Two main theories have been put forward concerning the interaction between life and environment. The idea about an old but almost static and uniform Earth with only slight gradual changes was defended by James Hutton (1726–1797) in his 'Theory of the Earth' and Charles Lyell (1797–1875) with the reputed 'Principles of Geology'.[30] The principle that 'the present is the key to understanding the past' is a logical consequence of these views. In this context, Charles Darwin (1809–1882) proposed the evolution of living beings by natural selection. In any case, his theory was based on gradual changes, without abrupt jumps: 'nature non facit saltum'.

These principles were challenged by the catastrophist theory, mainly developed by G. Cuvier (1769–1832) in his *Discours sur les Révolutions de la surface du*

[30] Four types of uniformities were the base of the theory: laws, processes (actualism), rates (gradualism) and state.

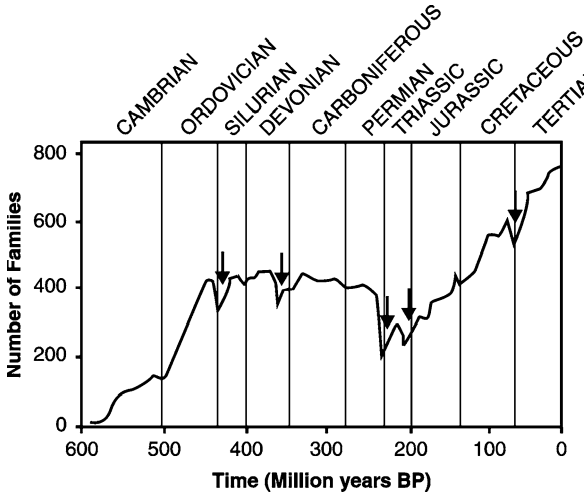

Fig. 2.36 Variation in the number of families along the Phanerozoic era. The major massive extinctions are marked with *arrows*. Reproduced from Steffen et al. (2005). Copyright: Springer Verlag

Globe. Based on geological observations of sediments, Cuvier assumed that biological evolution was driven by sudden events, producing the disappearance of some species and the emergence of new ones. Excluding astronomical causes, he favoured sudden changes in the positions of continents and oceans (see Rudwick 1997 for a translation of Cuvier's works with excellent commentaries on the source texts). According to this idea, processes operating in the past are not necessarily taking place today.

It soon becomes evident that the different biological species have rapidly evolved throughout time, some becoming extinct and others rapidly emerging. Alfred Wallace (1823–1913) expressed this with the words *We live in a zoologically impoverished world, from which all the hugest and fiercest and strangest forms have recently disappeared*. In short, the discussion was about the main driving agent in the biological evolution: chance (contingency) or necessity (natural selection), recalling Jacques Monod's (1917–1976) classic book 'Chance and Necessity'.[31]

The division of the history of the Earth in different periods (Table 2.1) has been marked, at least in the Phanerozoic era, by transitions coinciding with important changes in biological diversity. Living beings have a hierarchical classification: kingdoms, phyla, classes, orders, families, genera and species.

[31] In this context, the theory of 'punctuated equilibrium' should be mentioned, which supported the effect of chaotic events as a major force driving bursts in diversity. Evolutionary change occurs relatively rapidly in comparatively brief periods of environmental stress, separated by longer periods of evolutionary stability. This theory has been popularized in various books by S.J. Gould (1940–2002).

2.3.5.2 Biological Extinctions During the Phanerozoic Era

The record of the number of biological species reveals five important crises over the last 500 Ma, although many others are also evident (cf. Benton 1995; Hallan and Wignall 1997).

The extinction occurring at the end of the Ordovician period (443 Ma ago) was probably caused by a glaciation produced by the position of the supercontinent Gondwana, close to the South Pole (Sheehan 2001). Two pulses of extinction have been recorded, with the level of oceanic circulation playing a major role.

The end of the Devonian period was marked by an important decline in many biological species, the brachiopods and the foraminifera being specially affected (Wang et al. 1991 ; McGhee 1996). An impact at this time cannot be ruled out (Name 2003).

The end of the Permian period occurred 250 Ma before the present and was characterized by the extinction of 80% of all ocean-dwelling creatures and 70% of those on land (cf. Erwin 1994, 2006; Berner 2002; Benton 2003). Becker et al. (2001) discovery of fullerenes,[32] containing helium and argon with isotopic compositions similar to those in meteorites called carbonaceous chondrites, and the evidence that Becker et al. (2004) have recently found of an impact at the end of the Permian period in the crater Bedout (Australia) could point to the cause of this extinction.

At the boundary between the Triassic and Jurassic periods, about 200 million years ago, a mass extinction, occurring in a very short interval of time, destroyed at least half of the species on Earth (Ward et al. 2001). The thecodontians and many mammal-like reptiles became extinct, and this is the widely accepted view of how the dinosaurs attained dominance, as there were fewer predators to compete with them (Benton 1993). Olsen et al. (2002) claim to have found enhanced level of iridium at this stage. The Manicougan crater impact (210 $\pm$ 4 Ma old), located in Quebec (Canada), has been proposed as a possible scenario of the asteroidal impact.

However, the event which has been most studied is that which occurred at the end of the Cretaceous, no doubt because of its possible relation with the extinction of the dinosaurs.

2.3.5.3 The K/T Extinction

The Cretaceous extinction, hereafter called the K/T event,[33] is the one that has been most intensively studied. More than 75% of all the species present at that time became extinct, ranging across all families of organisms. De Laubenfels (1956) had already suggested that the extinction of the dinosaurs might have been caused by heat associated with the impact of a large meteorite. Urey (1973) and Hoyle and

[32] Large carbon compounds consisting of 60 or more carbon atoms, arranged as regular hexagons in a hollow shell. They are often called buckyballs, after Richard Buckminster Fuller (1895–1983), inventor of the geodesic dome, which their natural structure resembles.

[33] K stands for Cretaceous (from German) and T for the Tertiary Era – the Age of Mammals.

Wickramasinghe (1978) proposed, respectively, that ecodisasters such as the K/T extinction were probably caused by the collision or the close passage of a giant comet.

There was a surge of interest in the scientific and press media with the publication of the paper of Alvarez et al. (1980), which provided a degree of empirical association between massive biological extinction and the impact of a large (>10 km) extraterrestrial body, a comet or asteroid being the logical candidates (see Shoemaker 1983; Gehrels et al. 1994). Evidence of enhanced concentrations of iridium in the sediments corresponding to this period in the Italian site of Gubbio was verified by similar findings at other sites around the world (Orth et al. 1981). The publication of the Alvarez's paper immediately provoked different reactions supplying proof for and against this hypothesis.

After several failed attempts, the crater produced by the Cretaceous event was finally identified, the so-called 'smoking gun' being the Chicxulub crater on the Yucatan coast (Hillebrand et al. 1991). Kyte (1998) has tentatively identified materials that might have come from the impacting object. Figure 2.37 shows a view of the planet at this time.

These five large extinctions were not the only ones, and their magnitude is probably due to a combination of two or more external influences. The Earth mantle periodically undergoes instabilities, producing the eruption of the material to the surface in the form of giant volcanic events, called hotspots or superplumes

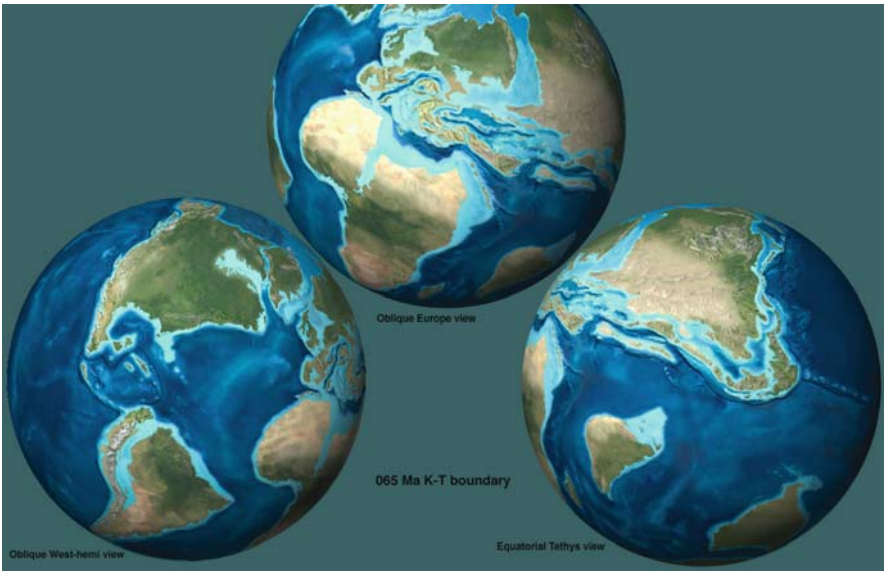

Fig. 2.37 Paleogeographic view of the Earth at the K/T transition. Courtesy: Ron Blakey, Department of Geology, Northern Arizona University

depending of their intensity and duration (Courtillot 1999; Condie 2001). The Permian (Siberian Traps) and Cretaceous (Deccan Traps) periods are associated with these manifestations, which lasted millions of years.

It has also been proposed (Abbott and Isley 2002) that large meteoric impacts could also trigger important volcanic eruptions. These authors proposed that the dating of the 38 known impact craters coincides with mantle eruptions. In principle, this could join the two main theories explaining the K/T mass extinction; however, the timing of mantle events, on the one hand, is not accurate enough and, on the other hand, they are characterized by quite different temporal extensions (see Palmer 1997). Purely biological explanations of mass extinctions should also be considered (see Raup 1992).

Physical conditions of the atmosphere have changed along the history of the planet and are likely to have been an important factor in the mass extinctions. At the geological scale, the climate is controlled by the balance between the variations in the solar output and the changes in the abundance of greenhouse gases in the atmosphere.

2.4 The Quaternary

The Quaternary Period is the geologic time period from the end of the Pliocene Epoch, roughly 1.8–1.6 million years ago to the present. The Quaternary includes two geologic subdivisions – the Pleistocene and the Holocene Epochs. This period is characterized by individual continents mainly accumulated in the Northern Hemisphere.

Significant growth of ice sheets did not begin in Greenland and North America until three million years ago, following the formation of the Isthmus of Panama by continental drift, thus preventing effective distribution of warm water in the North Atlantic Ocean. This triggered the start of a new era of rapidly cycling glacials and interglacials (Fig. 2.38 and Lisiecki and Raymo 2005).

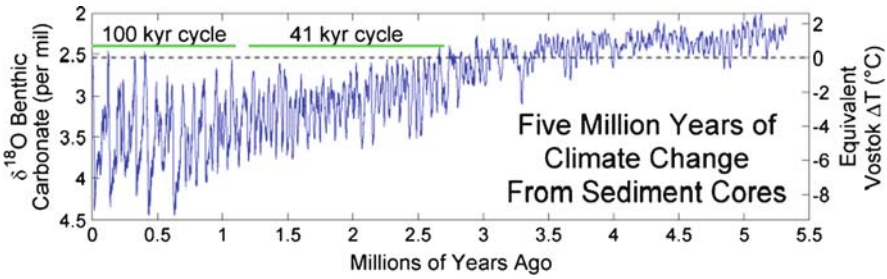

Fig. 2.38 Expanded view of climate change during the last five million years, showing the rapid oscillations in the glacial state. Present day is indicated by 0. Courtesy: Robert A. Rohde, Global Warming Art

2.4.1 The Ice Ages

The climate fluctuations in this period are mainly driven by variations in the orbital parameters of our planet, namely the inclination of the rotation axis, the eccentricity and the climate precession determining the time of the year where perihelion takes place. They are characterized by periods of 41,000, 100,000 and 19,000 years, respectively. In 1941, M. Milankovitch (1879–1958) calculated the insolation associated with these changes and proposed that the triggering of a glaciation was associated to the increase of ice-sheets in the northern hemisphere, produced at times of reduced temperature contrast between summer and winter.

Figure 2.39 shows the temperature record during the last half million years. The first two curves show local changes in temperature at two sites in Antarctica as derived from deuterium isotopic measurements on ice cores (EPICA Community Members 2004, Petit et al. 1999). The final plot shows a reconstruction of global ice volume based on measurements of oxygen isotopes on benthic foraminifera from a composite of globally distributed sediment cores and is scaled to match the scale of fluctuations in Antarctic temperatures (Lisiecki and Raymo 2005). The current decrease of the eccentricity of the Earth's orbit seems to delay a next glaciation, which may take place 50,000 years from now (Berger and Loutre 2002).

The temperatures oscillate in phase with CO_2 and CH_4 concentrations (Petit et al. 1999; Loulergue et al. 2008; Lüthi et al. 2008). Zeebe and Caldeira (2008) have shown that over the last 610,000 years the maximum imbalance between the

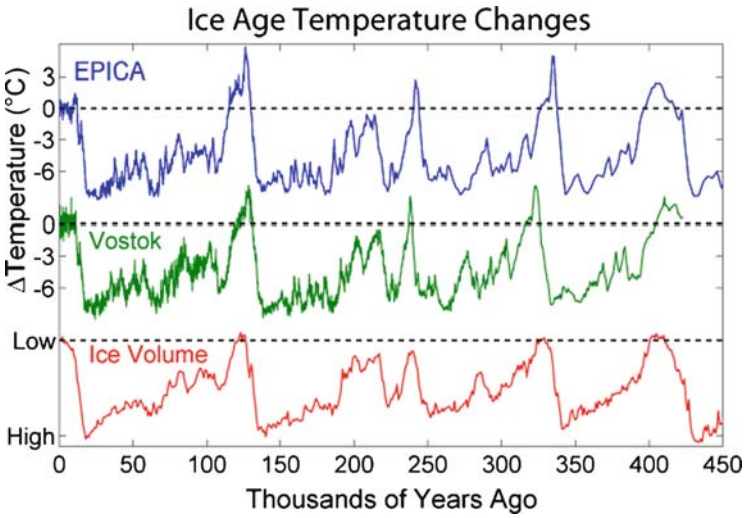

Fig. 2.39 Antarctic temperature changes during the last several glacial/interglacial cycles of the present ice age and a comparison to changes in global ice volume. The present day is on the *left*. Courtesy: Global Warming Project. Vostok data available at http://www.ngdc.noaa.gov/paleo/icecore/antarctica/vostok/vostok_data.html

supply and uptake of carbon dioxide (recall the phases of the CO_2 cycle) was 1–2% (~22 ppmv).[34] This long-term balance holds despite glacial–interglacial variations.

The 41,000 year periodicity, driven by obliquity changes, was dominant during the early Pleistocene. About 900,000 years ago this behaviour increased to a dominant 100,000 length of the glaciations linked to variation in the eccentricity. This is usually explained as an abrupt jump from one stable state of the climate system to another, typical of a non-linear response to a small external forcing (the insolation changes). Crowley and Hyde (2008) suggest that we are approaching to a new bifurcation, leading to a climate characterized by Antarctic-like 'permanent' ice sheets, which would shroud much of Canada, Europe and Asia. However, all these forecasts can be affected by the action of a new factor in the climate system: human intelligence.

2.4.2 The Present Warming: The Anthropocene

The last post-glacial era, known as the Holocene, was characterized by a stable climate (Fig. 2.40). During the last 7,000 years, the variations in the solar magnetic energy and the eruption of isolated volcanoes modulated the climate.

This period also signals the start of human civilization with the development of agriculture. Progressively, the activities of Homo sapiens became a significant force on the Earth (Burroughs 2005; Ruddiman 2005).

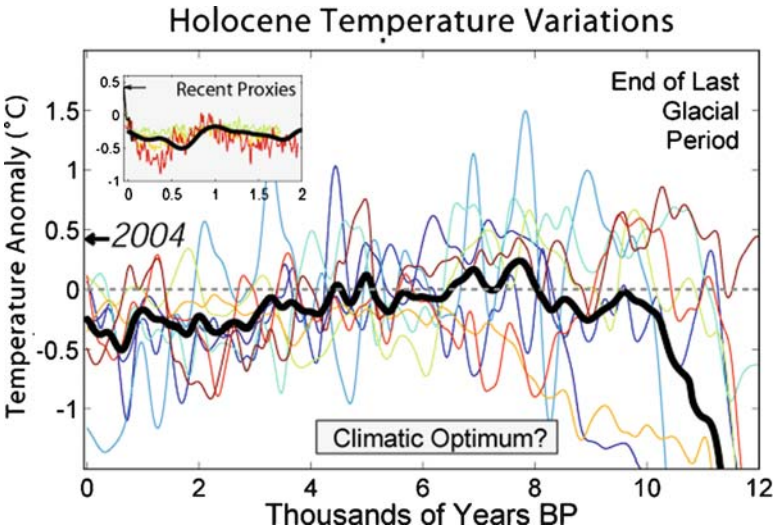

Fig. 2.40 Holocene temperature variations reconstructed from different sources. Courtesy: Robert Rohde

[34] ppmv: parts per million in volume.

We have previously seen how the living beings have clearly influenced the chemical composition of the atmosphere and the climate, through the emission of greenhouse gases resulting from their metabolism. At the time scale of a 1,000 years or less, the CO_2 cycle is dominated by two processes. One is when the gas is taken up via photosynthesis by green plants on the continents or phytoplankton in the oceans and later transferred to the soils or zooplankton, respectively; the plants, plankton and animals respire carbon dioxide and upon their death, they are decomposed by microorganisms with a subsequent production of CO_2. The other process is when carbon dioxide is exchanged between oceans and the atmosphere. As the cycle proceeds, concentrations of CO_2 can change as a result of perturbations in the cycle.

Probably the most characteristic event of the latter process started in the late 1700s, at which time the world entered the industrial era, which has continued into our days, an era in which many non-renewable resources are being used. The population of the planet exploded from 1 billion in 1850 to more than 6 billion by 2005, accompanied, for example by a growth in cattle population to 1,400 million (Mc Neill 2000). Crutzen and Stormer (2000) (see also Crutzen 2002a,b) named this period of the Earth history as the Anthropocene. Recently, Zalasiewicz et al. (2008) developed and updated this new concept.[35]

During this period the burning of fossil fuels produced large amounts of energy necessary for industry. As a consequence, residuals have been emitted to the atmosphere.

Georespiration is the principal process of atmospheric O_2 production at long time scales.

$$CO_2 + H_2O \longrightarrow CH_2O + O_2.$$

The opposite reaction is now occurring at a rate of about 100 times faster than that which would occur naturally.

$$CH_2O + O_2 \longrightarrow CO_2 + H_2O.$$

As a result, the long-term carbon cycle impinges on the short-time cycle, leading to an extremely fast rise in atmospheric CO_2 (Fig. 2.41). The CO_2 cycle cannot balance the disequilibrium produced by humans.

Because the carbon dioxide emitted in the combustion of fossil fuels is a greenhouse gas, we can expect a rise in mean temperatures. Figure 2.42 shows the increase of globally averaged temperatures during the instrumental record. The future warming will depend on future emissions of greenhouse gases. The recent IPCC report, 2007, established different scenarios with temperature increases in the range 2–5°C for 2100.

[35] Written by 21 members of a commission organized by the London Geological Society to elucidate if we have entered a new geological period. This commission unanimously answered 'yes' to the question: 'Are we now living in the Anthropocene?'.

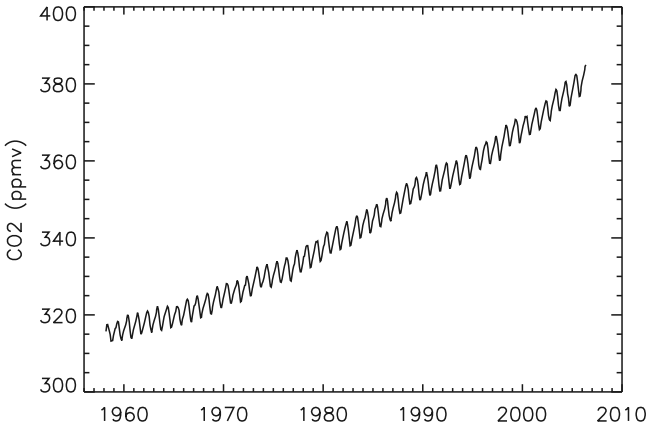

Fig. 2.41 Variation of the atmospheric concentration of carbon dioxide, as measured in Mauna Loa (Hawaii). Data: Earth System Research Laboratory (NOAA)

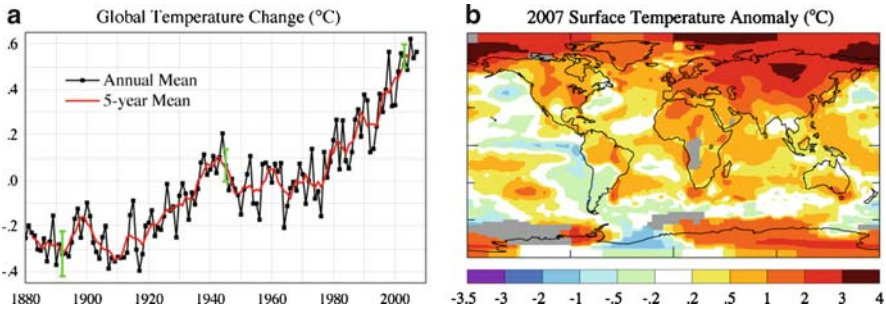

Fig. 2.42 (*Left*) Global annual surface temperatures relative to 1951–1980 (mean figures), based on surface air measurements at meteorological stations and ship and satellite measurements for sea surface temperature. (*Right*) Colour map of temperature anomalies in 2007 relative to the 1951–1980 mean. Areas that were warmest in 2007 are in *red*, and areas that have cooled are in *blue*. Courtesy: Goddard Institute for Space Studies

We may be at the beginning of a sixth massive extinction (Thomas 2004). During the mentioned Anthropocene period, only 40 animals and around 100 plants increased their numbers.[36]

The impacts of current human activities will continue over long periods and the climate may depart significantly from natural behaviour over the next thousands of years. We should remark that the Earth System has critical processes that are susceptible to abrupt changes triggered by human activities that will render the planet less hospitable for human life. The climate system, in particular the sea level, may be responding more quickly to climate change than the current generation of models indicates (Rahmstorf et al. 2007).

[36] They include humans, domesticated plants and animals and synanthropes (e.g. rats, rabbits etc.).

For further information on this topic, see the following monographs: Houghton 1997, 2005 and Stern 2006, and the successive IPCC reports[37] Houghton et al. 1996, 2001, IPCC 2007.

Referring to this topic, Carl Sagan commented in his book 'Pale Blue Dot': *Some planetary civilizations see their way through, place limits on what may and what must not be done, and safely pass through the time of perils. Others, not so lucky or so prudent, perish.*

2.5 The Future of Earth

We have already seen how the presence of life has produced a cooler Earth. A resulting feedback of this connection has been the extension of our estimates on the lifespan of the biosphere (Lenton and Van Bloh 2001). However, this process will also have a limit.

2.5.1 The End of Life

The long-term future of our planet will be driven by the continuous increase of solar luminosity (Fig. 2.43a). In response, the weathering rate will increase, lowering rapidly the concentration of CO_2 in the atmosphere and soil (Fig. 2.43b). At a given time both will reach minimum values for the survival of plants (10 ppm).

The disappearance of life will follow the reverse sequence to its appearance in our planet. Franck et al. (2005) have predicted that the extinction of multicellular life and eucaryotes will take place at 0.8 and 1.3 Ga from now, respectively. For earlier estimations of the biosphere life span, see Lovelock and Whitfield (1982) and Caldeira and Kasting (1992). Therefore, it seems that the time period in the Earth's history in which complex life exists is very short. The consequences of this process for the perspectives of finding alien complex life have recently been popularized by Brownlee and Ward (2000, 2004), with the 'Rare Earth' hypothesis. For critical overviews on this topic see Kasting (2000) and Darling (2002).

Two events will make the situation much more complicated. As a result of increasing temperatures, the Earth will enter a runaway greenhouse phase, similar to that suffered by Venus, probably 1.2 Ga from now; as a first consequence, the oceans will evapourate. Additionally, tectonic activity will be interrupted due to the progressive exhaustion of internal energy sources.

The end of life described is limited to photosynthetic organisms. Organisms that use chemical energy for their metabolism could probably survive, especially underground.

[37] Since its founding, in 1988, the Intergovernmental Panel on Climate Change (IPCC) has published four reports: 1990, 1995, 2001 and 2007.

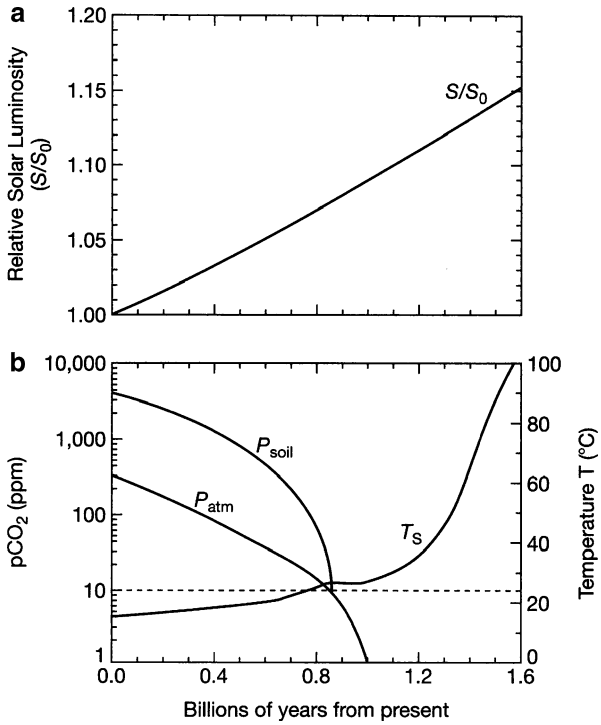

Fig. 2.43 Long term projections of (**a**) solar luminosity and (**b**) surface temperature and CO_2 concentration in the soil and in the atmosphere. Adapted from Caldeira and Kasting (1992) Fig. 2. Reprinted by permission from Macmillan Publishers Ltd. Nature Vol. 360 p. 721. Copyright (1992)

2.5.2 The End of the Earth

About 5.5 Ga from now, a significant amount of the hydrogen reserves of the Sun will have been spent, and our star will move to the Red Giant Branch (RGB) phase of its evolution. The solar diameter will expand to roughly 150 times its current diameter, reaching 0.77 AU, while the core will start to collapse under its own weight, getting hotter and denser. This phase will end with the ignition of Helium in the core, but this element is consumed rapidly and the star will contract again, leading to a second expansion of the outer layers (Asymptotic Giant Branch, AGB).

Mercury and Venus will clearly be affected during the RGB phase. The strong mass loss during this period will expand the Earth's orbit, probably to 1.6 AU, saving it temporarily from being destroyed. This will inevitably occur during the AGB phase, when the Sun undergoes a second expansion (see Rasio et al. 1996 for a consideration of the tidal decay in the final stage). The Earth will probably be vapourized, and the remains scattered through the interplanetary medium (Sackmann et al. 1993; Schröder et al. 2001 and Rybicki and Denis 2001). Recently, Schröder and Smith (2008) have presented a new RGB model, concluding that the

Earth's engulfment will happen during the late phase of the solar RGB evolution. The minimum orbital radius for a planet to be able to survive is found to be about 1.15 AU.

Villaver and Livio (2007) studied the survival of gas planets around stars with masses in the range 1–5 M_S, as these stars evolve off the main sequence. They show that planets with masses smaller than one Jupiter mass do not survive the planetary nebula phase if located initially at orbital distances smaller than 3–5 AU.

Retter and Marom (2003) interpreted an outburst, with at least three peaks, of the red giant star V838 Monocerotis to have been produced by the engulfment of three nearby planets. However, the death sentence of our planet is not yet final and two observations bring some hope. Silvotti et al. (2007) have detected at 1.7 AU a giant planet orbiting the red giant V391 Pegasi, which is already burning helium in its core, whereas Mullally et al. (2008) suggest the presence of a giant planet (2.1 Jupiter masses) orbiting a white dwarf at 2.36 AU. Moreover, we cannot forget that the first exoplanets were found orbiting the pulsar PSR1257+12, enduring a supernova explosion (Wolszczan and Frail 1992).

In any case, in a long-term perspective the human race must move into space to avoid extinction. Following the famous quote of K. Tsiolkovsky (1857–1935): *Earth is the cradle of humanity, but one cannot remain in the cradle forever.*

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