The first 660 million years of the Earth’s existence, known as the **Hadean**, was the stage during which the metallic core separated from the silicate mantle, the atmosphere and hydrosphere were formed, and melting of the silicate mantle produced the earliest crust.

The early Earth was violent and hot, and giant impacts would have been devastating. The heat released would have been capable of melting the outer part of the Earth to form a **global magma ocean**.

virtually no direct rock record

theoretical modelling and geochemistry to reveal the mechanisms of formation of the different layers in the Earth and the timescales involved.

Evidence for the development of the early Earth-

(1) Reviews the various heat sources necessary to drive planetary differentiation.

(2) Investigates

-the mechanisms of core formation,

-the evidence for the presence of a magma ocean,

-and the timing of accretion and core formation.

(3) Explores the evidence for the origin and age of the Moon.

(4) Formation and evolution of the atmosphere and hydrosphere, and finally, the evidence for the nature and formation of the earliest continental crust will be reviewed.

You will discover that the processes that formed the layering in the early Earth, atmosphere and hydrosphere effectively shaped the planet as we know it today,

Heating and differentiation of the Earth

$\boxed{\text{Differentiation is the process by which planets develop concentric layering, with zones that differ in their chemical and mineralogical compositions.}}$

The generation of such zones results from a differential mobility of elements due to differences in their physical and chemical properties.

When a rock is heated, different minerals within the rock will melt at different temperatures. This phenomenon is known as **partial melting** and is a key process in the formation of liquid rock or magma,

Once elements have been mobilized in this manner, they will begin to migrate under the influence of pressure or gravity.

Imagine a body the size of a planetary embryo that had accreted from nickel–iron and silicate minerals. Nickel–iron has a density of about 7.9×10^3 kg m^{-3} (compared to ~3.0×10^3 kg m^{-3} for silicate minerals) and a melting point some hundreds of degrees higher than silicates. What would happen if temperatures within this planetary embryo were increased to a point at which silicates began to melt?

Since nickel–iron has a higher melting point, it would remain solid after the silicates had begun to melt and, because it is much denser than any silicate minerals, it would begin to sink towards the centre of the body.

In addition to the separation of a metallic core from a silicate mantle, partial melting of planetary bodies also produces residual solid silicate minerals in contact with a silicate melt that have different compositions, depending on how extensive the melting is.

In such a system, incompatible elements are partitioned into the melt more readily than compatible elements **(element partitioning)**, is the principal mechanism by which incompatible elements first become concentrated into the melt.

Then, if the magma is buoyant, they migrate upwards to form the overlying crust.

Heat sources

If partial melting is the principal cause of differentiation, then the Earth needs to be heated before layering can begin to develop. There are several sources of heat that can arise during Earth's evolution.

The most important are:

- Primordial heat sources, which develop in the early stages of planetary evolution (i.e. those associated with accretion, collision and core formation)
- Tidal and radiogenic heating processes, which can operate long after the planet has been formed.

Upon hitting the Earth, if all the kinetic energy of motion is converted into heat, then the increase in temperature, ΔT, can be calculated:

$$\Delta T = mv^2/2(m + M)C$$

where the body (of mass m) impacts the Earth (of mass M) and C is the specific heat capacity of Earth material (i.e. the amount of heat required to raise the temperature of 1 kg of material through 1 K).

Not all of the impacting material arrived at the same time: accretion took place over ~10^7 years. More importantly, not all of the kinetic energy would be converted to heat.

For example, some of the impact energy would be used in
1. the excavation of large craters
2. much of the heat would have been radiated into space.

Nevertheless, most estimates predict temperatures to have risen above the melting point of silicate minerals and iron–nickel, which means that the Earth is likely to have gone through an early molten stage.

1. A planetesimal of mass 10^{15} Kg impacts the Earth with a velocity of 10,000 m/second. Calculate the rise in temperature in the Earth assuming that the heat generated by the impact spreads rapidly and uniformly throughout the whole Earth. Because m is much smaller than M, the effect of m is negligible and can be ignored, so equation can be simplified to $\underline{\Delta T = mv^2/2MC}$
Take the total mass of the Earth to be $\underline{6 \times 10^{24}\ Kg}$ and the average specific heat capacity of the Earth to be $\underline{750\ J\ Kg^{-1}\ K^{-1}}$ (note: $1\ J = 1\ Kg\ m^2\ s^{-2}$)
2. Suppose that the Earth was constructed entirely of 10^{15} Kg planetesimals, each of which generated the temperature rise obtained in question 1. What would be the total temperature rise?

Answers to Questions no. 1 and no.2-

1. From Equation $\Delta T = mv^2 / 2MC$

Where m= 10^{15} Kg (given), v = 10^4 m/s (given), C= 7.5×10^2 J/Kg-K (given) and M = 6.0×10^{24} Kg (given).

$\Delta T = 10^{15}\ Kg \times (10^4\ m\ s^{-1})^2\ /\ 2 \times (6.0 \times 10^{24}\ Kg) \times (7.5 \times 10^2\ J\ Kg^{-1}\ K^{-1}) = \underline{1.1 \times 10^{-5}\ K}$

2. The number of planetesimals of mass 10^{15} Kg each required to construct the Earth is:

$6 \times 10^{24}\ Kg / 10^{15}\ Kg = 6 \times 10^9$

If each of these produced the temperature rise in part (a), the total temperature rise in the Earth would be $(6 \times 10^9) \times (1.1 \times 10^{-5}\ K) = 66000$ K.

Core formation

If the Earth went through an early molten phase, allowing the metals and silicates to separate, then the 'falling inwards' of the nickel–iron-rich fraction to form the core would have released potential energy.

The gravitational energy lost by the inward movement of nickel–iron would have been converted first to kinetic energy and then into thermal energy.

It is estimated that the core-forming process would have contributed significantly to the Earth's primordial heating (though it would still have been an order of magnitude less than that generated by collision and accretion).

if these primordial heat sources had remained the only way of heating the Earth, their intensity would have waned through time due to continual radiative heat loss to space.

Since heat drives fundamental processes such as volcanism, the fact that the Earth has remained volcanically active to the present day requires additional processes of internal heat generation.

Radiogenic heating

During the latter half of the 19th century, the eminent physicist Lord Kelvin (1824–1907) attempted to determine the age of the Earth. He believed the Earth had cooled slowly after its formation from a molten body and assumed the main sources of energy were from primordial heat and tidal friction.

Taking many factors into account, including the mass of the Earth, the current rate of surface heat loss, and the melting points of various rock types, he concluded that the planet could not be much older than about 20–40 Ma.

Kelvin’s calculations did not gain wide acceptance with geologists he was a powerful scientific influence of the time, and it was not until much later in the 1950s that accurate radiometric dating experiments eventually proved him wrong.

These experiments were conducted in meteorites and demonstrated that the formation of the Earth occurred about 4.6 Ga ago

Table 2.1 Naturally occurring radioactive decay systems of geochemical and cosmochemical interest.

Parent	Decay mode*	Decay constant (λ)/ y^{-1}	Half-life/y	Present heat production‡	Daughter isotopes	Measured ratio
^{40}K	β+, e.c., β–	5.54×10^{-10}	1.28×10^{9}	2.8	^{40}Ar, ^{40}Ca	$^{40}Ar/^{36}Ar$
^{87}Rb	β–	1.42×10^{-11}	4.88×10^{10}		^{87}Sr	$^{87}Sr/^{86}Sr$
^{147}Sm	α †	6.54×10^{-12}	1.06×10^{11}		^{143}Nd	$^{143}Nd/^{144}Nd$
^{187}Re	β	1.59×10^{-11}	4.35×10^{10}		^{187}Os	$^{187}Os/^{188}Os$
^{232}Th	α	4.95×10^{-11}	1.39×10^{10}	1.04	^{208}Pb, ^{4}He	$^{208}Pb/^{204}Pb$, $^{3}He/^{4}He$
^{235}U	α	9.85×10^{-10}	7.07×10^{8}	0.04	^{207}Pb, ^{4}He	$^{207}Pb/^{204}Pb$, $^{3}He/^{4}He$
^{238}U	α	1.55×10^{-10}	4.47×10^{9}	0.96	^{206}Pb, ^{4}He	$^{206}Pb/^{204}Pb$, $^{3}He/^{4}He$
^{26}Al	β–	9.5×10^{-7}	0.73×10^{6}		^{26}Mg	$^{26}Mg/^{24}Mg$
^{129}I	β–	4.41×10^{-8}	1.57×10^{7}		^{129}Xe	$^{129}Xe/^{130}Xe$
^{146}Sm	α	6.73×10^{-9}	1.03×10^{8}		^{142}Nd	$^{142}Nd/^{144}Nd$
^{182}Hf	β–	7.78×10^{-8}	8.9×10^{6}		^{182}W	$^{182}W/^{184}W$
^{244}Pu	α, SF	8.45×10^{-9}	82×10^{6}		^{n}Xe	$^{n}Xe/^{130}Xe$**

*α = alpha decay (^{4}He); β^{-} = beta decay (electron or positron); e.c. is electron capture; SF is spontaneous fission.

†The production of ^{4}He from ^{147}Sm decay is insignificant compared with that produced by decay of U and Th.

‡Heat production averaged for the whole Earth in units of 10^{-12} W kg^{-1} of Earth material (not of the isotope).

**n can be 124, 126, 128 or 129, all of which are produced by ^{244}Pu fission. Element symbols are listed in the Appendix.

Which of the isotopes remain active today and which are extinct?

All those with half-lives significantly less than the age of the Earth, i.e. 4.6 Ga, are extinct, namely: 26Al, 129I, 146Sm, 182Hf and 244Pu.

The others, principally isotopes of 40K, 87Rb, 147Sm, 232Th, 235U and 238U, are still active today.

If radionuclides have short half-lives and are not replenished by the decay of other isotopes, then they may be lost altogether. One such short-lived extinct nuclide is 26Al, which has a half-life of 0.73 Ma.

What evidence would you look for to support the presence of 26Al in the early Solar System?

26Al decays to 26Mg, so you would expect to see anomalously high abundances of 26Mg relative to other isotopes of Mg in materials from the early Solar System.

Studies of primitive meteorites, notably carbonaceous chondrites, do indeed show slightly high 26Mg/24Mg ratios, which suggests that a significant proportion of the aluminium present at the time of condensation of the solar nebula was the unstable isotope 26Al.

What does the observation that 26Al was present in chondritic meteorites tell you about the timescale of the formation of the Solar System?

The half-life of 26Al is only 0.73 Ma, so the time between the supernova explosion that generated the 26Al and the accretion of the meteorite parent body must have happened on a similar timescale of a few million years.

Given that after 10 half-lives only 1/2^10 (or1/1024 th) of the original number of 26Al atoms remain, then for any measurable amount of radiogenic 26Mg to be found, chondritic meteorites must have formed within, at most, **7.3 Ma of the supernova**.

Whilst such short-lived isotopes may have been important heat sources during the early stages of terrestrial planet evolution, study of Earth material indicates that it is the isotopes of the elements U, Th and K that are responsible for most of the radiogenic heating that has occurred throughout the history of the planet.

These isotopes, which all have particularly long half-lives, are termed long-lived radiogenic nuclides and were present in sufficient quantities after condensation and accretion to ensure that they have remained abundant within present-day Earth.

U, Th and K are incompatible elements (i.e. they are preferentially partitioned into silicate liquids), so where will they be concentrated in the Earth?

The elements U, Th and K (and their radiogenic isotopes) are particularly concentrated in the silicate-dominated outer layers of the Earth, and in particular within the continental crust. They are thought to be virtually absent from the core.

Finally, whilst the rate of radiogenic decay is constant for each isotope system, the total amount of radioactive decay, and hence heat generation, will decline over time as the reserves of the original radioactive materials are gradually used up. This gradual depletion of radioactive materials is expressed in half-lives, and each isotopic decay system has its own unique half-life.

For example 235U
decays through a series of α-particle emissions to the daughter isotope 207Pb, after 7.07 × 10^8 years, half of the 235U originally present will have decayed to 207Pb and the remainder will continue to halve every 7.07 × 10^8 years. Over the age of the Earth (4.6 Ga), approximately 6.5 half lives of 235U have elapsed, so the heat production from 235U is now (½)6.5 ≈ 0.01 (i.e. 1%) of what it was originally.