# Origin of the Core

Francis Nimmo

Dept. Earth Sciences, University of California, Santa Cruz

F. Nimmo, Dept. Earth Sciences, University of California, Santa Cruz, CA 95064, USA

(fnimmo@es.ucsc.edu), tel. 831-459-1783, fax. 831-459-3074

## Origin of the Core

All major bodies of the inner solar system, including the Earth, are composed primarily of iron and silicates. Due to its high density, the iron component - the core - is invariably found at the centre of the body. In addressing the origin of the Earth's core, four main questions arise. Firstly, composition: of what does the core consist? Secondly, accretion: how and when was this material assembled? Thirdly, differentiation: how and when did the Earth develop a recognizable core at its centre? And finally, evolution: how did the core change subsequent to its formation? The first of these questions is dealt with elsewhere (see Core Composition; Inner core: composition), and will be only briefly summarized below. The other questions form the bulk of this article. Although the focus of the article will be the Earth's core, where possible these questions will also be discussed for other terrestrial planets. The answers to these questions are in many cases poorly understood, especially for planets for which few or no samples are currently available.

### 1) Composition

Because the solar photosphere and primitive (chondritic) meteorites have similar ratios of most elements, it is a reasonable supposition that these elements are present in the same ratios in the bulk Earth (and other terrestrial planets) (Taylor, 1992). The chondritic Si:Fe atomic ratio of 1.2:1 suggests that terrestrial planets should have roughly comparable core and mantle masses. The actual ratios of core to mantle mass are 85:15, 33:67, 15:85 and 2:98 for Mercury, Earth/Venus, Mars and the Moon (Lodders & Fegley, 1998). This rather wide range of values is probably a result of late-stage impacts, as discussed below.

The composition of planetary cores depends partly on what elements were present at the time of core formation. Simple models of solar system formation suggest that elements with relatively low condensation temperatures (i.e. volatiles) are likely to have been more abundant at greater distances from the Sun. The terrestrial planets and meteorites appear to be depleted in elements with condensation temperatures $<\sim 1000$ K, compared to the initial solar nebula (Taylor, 1992). Whether there are gradations in volatile content between the terrestrial planets is currently unclear. Potassium isotope data (Humayun & Clayton, 1995) are a strong argument against such gradients occurring as a result of fractionating processes. Furthermore, any initial gradation will have been reduced by mixing of bodies from different solar distances during the accretion process.

For the Earth, more information is available because of seismological constraints on its interior structure, and constraints from samples on its bulk composition. As discussed elsewhere (Core composition; Inner core: composition), the core consists of Fe plus about 10 wt% of one or more light element(s), probably either S,Si or O. These light elements are more concentrated in the liquid outer core, and can significantly reduce the melting temperature of the liquid (see Melting temperature of iron in the core - theory; Ideal Solution theory). If core solidification occurs at the centre of the planet, expulsion of the light elements during freezing can drive compositional convection (see Convection, chemical; Energy source for the geodynamo).

Because alkali element abundances are depleted with respect to the solar nebula, it is likely that potassium was mostly lost as a volatile during Earth's formation. Nonetheless, the remaining potassium is a potentially significant constituent of the core. Recent experimental evidence and theoretical arguments both suggest that potassium may partition into the core, especially if sulphur is a major constituent (see Radioactive isotopes, their decay in core & mantle). Radioactive decay of potassium-40 may help to power the terrestrial geodynamo (see Energy source for the dynamo; Nimmo et al. 2004), and could also be important on other planets. The presence of these secondary elements is presumably determined by their overall availability, and the conditions which applied the last time the iron was in equilibrium with its surroundings.

There are few constraints on the detailed compositions or states of other planetary cores (see Taylor (1992) and Nimmo & Alfe (2006) for summaries). The Moon is very depleted in volatile elements and has a small core; these observations can be explained if it accreted at high temperatures from debris left over after a Mars-sized object collided with the Earth (e.g. Canup & Asphaug, 2001). Analyses of meteorites thought to have come from Mars have been used to deduce a planet that is richer in S and Fe than the Earth. The response of Mars to solar tides has been used to argue that the core of Mars is at least partially liquid. Similar observations have also been made for the Moon, Mercury and Venus. Since most of these bodies are small and cool rapidly, these observations strongly suggest the presence of substantial amounts ($\sim$10 wt%) of S in the outer cores of these bodies. Because S is not incorporated into the solid core at low pressures, it becomes progressively harder to freeze the remaining liquid as solidification progresses. Whether the presence of large quantities of S is likely at Mercury's orbit, or following a high energy impact, remains an unanswered question.

**2) Solar System Accretion**

The process of accretion is now reasonably well understood, at least in outline (see Weidenschilling (2000) and Chambers (2004) for summaries of the likely timescales and processes). The solar system began when a nebular cloud of dust and gas became gravitationally unstable, and collapsed in on itself, forming a star surrounded by the orbiting remnants of the original cloud. Models suggest that the accumulation of these cloud remnants into Mars-sized planetary embryos was a rapid process, taking O($10^5$ years) in the inner solar system and somewhat longer at greater distances from the Sun. The final stage of accretion, in which these planetary embryos collided in high-energy impacts, was much slower, taking O($10^8$ years). The early formation of Jupiter is presumed to have frustrated planet formation within the asteroid belt (see Chambers, 2004).

Presumably, each planetary embryo began as a relatively homogeneous object, similar to the undifferentiated asteroids observed today. However, collisions between different embryos caused these bodies to increase in temperature as they grew (e.g. Stevenson, 1989). Furthermore, bodies which formed early may also have been heated by the decay of now-extinct radioactive elements, especially $^{26}$Al (half-life 0.73 Myr). Thus, at some critical size, the planetary embryo will have begun to melt and (as discussed below) core formation is likely to have proceeded rapidly thereafter. This critical size probably varied in both time and space; for instance, Ceres, the largest asteroid (present-day diameter D=913 km) appears not to have differentiated at all, while Vesta (D=520 km) underwent widespread melting which certainly involved core formation.

The process of differentiation is discussed in more detail below. Of more interest, from the point of view of accretion, is the timescale over which differentiation occurs. This timescale may be obtained from hafnium-tungsten (Hf/W) isotopic observations, as follows (see Harper & Jacobsen (1996) for a summary).

Core differentiation leads to fractionation between Hf and W, with Hf going preferentially into the silicate phase, and about 80% of the W into the core. The unstable isotope of Hf, $^{182}$Hf, decays to $^{182}$W with a half-life of 9 Myrs. If differentiation happens before all the $^{182}$Hf is exhausted, then the mantle will show an excess of $^{182}$W over stable $^{180}$W relative to an undifferentiated (chondritic) sample. The amount of the $^{182}$W excess depends on the initial $^{182}$Hf/$^{180}$Hf ratio and the timing of differentiation. Since the former is known, observations of the $^{182}$W/$^{180}$W ratio have been used to constrain core formation timescales.

If core formation is assumed to occur instantaneously, then the times after solar system formation at which core formation took place are (Kleine et al. 2002): 4 Myr,13 Myr,33 Myr and 26-33 Myr for Vesta, Mars, the Earth and the Moon, respectively. These timescales indicate that planetary accretion and differentiation is a relatively rapid process. Core formation, as discussed below, occurs most rapidly if the iron is liquid. Vesta's early core formation age and small size suggest that heating either by $^{26}$Al or large impacts was probably important. For larger bodies like the Earth, core melting would have occurred simply due to the gravitational energy of accretion (Stevenson, 1989; see below).

In practice, accretion is likely to occur over a long timescale (tens of Myrs for an Earth-sized body; Chambers, 2004). Furthermore, for bodies the size of the Earth, it is likely that many of the accreting planetary embryos had themselves already undergone differentiation. Thus, the interpretation of the apparent core age is not straightforward. For instance, the increase in apparent core age with object size (Fig 1) is probably a reflection of a more prolonged accretion process for the larger bodies. An additional problem in the interpretation of Hf/W data is that the degree to which incoming bodies re-equilbrate with the planet's mantle is uncertain, and probably depends on the size of the impactor and the state of the mantle (magma ocean or solid) (Harper & Jacobsen 1996). The Moon-forming impact almost certainly produced a magma ocean on Earth, and may also have partially or completely re-set the isotopic Hf/W clock.

Similar issues arise when considering the concentrations of siderophile (iron-loving) elements in the mantle (see Righter (2003) for a review). On Earth, these concentrations are significantly higher than those expected for equilibration at low pressures and temperatures. A possible solution is that equilibration took place at the bottom of a deep ($\sim$1000 km) magma ocean. Interestingly, siderophile abundances estimated for the Martian mantle, based on Martian meteorite compositions, are more consistent with equilibration at lower temperatures and pressures (Righter, 2003). Although highly model-dependent, these observations suggest that Mars lacked a deep magma ocean. These conclusions are consistent with the expectation that larger bodies have higher internal temperatures owing to their greater gravitational potential energy.

**3) Differentiation**

As described above, small planetary embryos (or planetesimals) probably consisted of a homogeneous mixture of metallic, silicate and volatile components. However, as the planetesimals grew, collision velocities and thus the heat energy of each impact increased. Simultaneously, radioactive decay within the planetesimals further increased their internal temperature, especially if short-lived species such as $^{26}$Al were still active at this point. At some particular size, probably depending on the rate of growth, the interior of the planetesimal started to melt. As described below, melting permits the denser components of the planetesimal to migrate towards the centre of the body, thus achieving the process of differentiation. The following description is based largely on Stevenson (1990) and Solomon (1979) (see also Fig 2).

In general, metallic iron melts before end-member mantle silicates. Thus, differentiation can potentially occur either by the core liquid migrating through a solid silicate matrix, or by liquid core droplets settling in a silicate liquid. The latter process is appropriate e.g. to the kind of magma ocean that is formed after giant impacts, and results in rapid core formation. Based on the balance between viscous and surface tension forces in a convecting magma ocean, the likely iron droplet size is $\sim$1 cm. At these length-scales, chemical equilibration with the surrounding mantle material is likely to be complete.

The former process, core liquid migrating through a solid silicate matrix, is controlled largely by the dihedral angle, the characteristic angle formed at an interface between solid grains and an interstitial melt. For large dihedral angles, typical of iron-silicate interfaces at low pressures, melt percolation is inefficient because the iron droplets accumulate in disconnected pockets. Percolation is much more rapid if the dihedral angle is smaller ($<60^\circ$), the melt fraction is large ($>\sim10\%$), or large shear stresses occur. The dihedral angle typically decreases with pressure, so that large variations in permeability with depth may have occurred within accreting planets. For dihedral angles $<60^\circ$, transit timescales through the mantle for individual iron particles are typically $10^4 - 10^5$ yrs. For dihedral angles $>60^\circ$, an alternative mechanism for differentiation is the accumulation of large bodies of iron which then migrate downwards as diapirs.

Once differentiation begins, it is likely to go rapidly to completion: the downwards motion of the dense iron results in a release of gravitational potential energy, leading to local heating, viscosity reduction and an increase in the rate at which differentiation proceeds. This runaway effect is larger for larger bodies, but even for Mars-sized objects the energy of core differentiation is sufficient to raise the temperature of the entire planet by 300 K (Solomon, 1979).

**4) Evolution and inner core growth**

Once the bulk of differentiation has ended, subsequent evolution of planetary cores is more leisurely. Core cooling is controlled by the rate at which the overlying mantle can remove heat (see Heat flow across the core-mantle boundary; Interiors of planets and satellites). Reactions between the core and mantle may take place, but are likely to be relatively slow ( see D", Composition). Partly because of the energy released during differentiation, the cores are initially likely to be both hot and convecting, and in many cases (e.g. Mars, Earth, the Moon?) develop early dynamos (see Dynamos in Planets & Satellites; Magnetic Field of Mars; Nimmo & Alfe, 2006).

Over the longer term, the core will cool and core solidification will set in, acting as another energy source for the dynamo (see Energy source for the geodynamo). Only for the Earth is the existence of a solid inner core certain (see Inner core: composition). Similar inner cores may exist for the other terrestrial planets, depending on the slopes of the core adiabat (see Adibatic gradient in the core) and melting curve, and the core sulphur content (Nimmo & Alfe, 2006). The age of the inner core depends on the rate at which the mantle is extracting heat from the core. Current estimates (see Heat flow across the core-mantle boundary) suggest an inner core age of only $\sim$1 Gyr. Although the uncertainties in this value are still quite large, a cooling rate sufficient to maintain the terrestrial dynamo implies an inner core significantly younger than the age of the Earth (Nimmo et al. 2004). Prior to the formation of the inner core, the terrestrial dynamo was driven mainly by cooling of the liquid core, with possible assistance from radioactive decay (see Radioactive isotopes, their decay in core & mantle).

**References**

Canup, R.M. and E. Asphaug, Origin of the Moon in a giant impact near the end of the Earth's formation, Nature 412, 708-712, 2001.

Chambers, J.E., Planetary accretion in the inner Solar System, Earth Planet. Sci. Lett. 223, 241-252, 2004.

Harper, C.L. and S.B. Jacobsen, Evidence for 182Hf in the early solar system and constraints on the timescale of terrestrial accretion and core formation, Geochim. Cosmochim. Acta 60, 1131-1153, 1996.

Humayun, M. and R.N. Clayton, Potassium isotope cosmochemistry - genetic implications of volatile element depletion, Geochm. Cosmochim. Acta 59, 2131-2148, 1995.

Kleine, T., C. Munker, K. Mezger and H. Palme, Rapid accretion and early core formation on asteroids and the terrestrial planets from Hf-W chronometry, Nature 418, 952-955, 2002.

Lodders, K. and B. Fegley, The planetary scientist's companion, Oxford Univ. Press, Oxford , 1998.

Nimmo, F. and D. Alfe, Properties and evolution of the Earth's core and geodynamo, in Advances in Science: Earth Science, P.R. Sammonds and J.M.T. Thompson, eds., Imperial College Press, 2006.

Nimmo, F., G.D. Price, J. Brodholt, D. Gubbins, The influence of potassium on core and geodynamo evolution, Geophys. J. Int. 156, 363-376, 2004.

Righter, K., Metal-silicate partitioning of siderophile elements and core formation in the early Earth, Ann. Rev. Earth Planet. Sci. 31, 135-174, 2003.

Solomon, S.C., Fomation, history and energetics of cores in the terrestrial planets, Phys. Earth Planet. Inter. 19, 168-182, 1979.

Stevenson, D.J., Formation and early evolution of the Earth, in Mantle convection and plate tectonics, W.R. Peltier, ed., pp. 818-868, Gordon and Breach, 1989.

Stevenson, D.J., Fluid dynamics of core formation, in Origin of the Earth, eds. H.E. Newsom and J.E. Jones, pp. 231-249, Oxford Univ. Press, New York, 1990.

Taylor, S.R., Solar system evolution, Cambridge Univ. Press, Cambridge, 1992.

Weidenschilling, S.J., Formation of planetesimals and accretion of the terrestrial planets, Space Sci. Rev. 92, 295-310, 2000.

**Cross-references**

Adiabatic gradient in the core

Convection, chemical

Core Composition

D”, Composition

Dynamos in Planets & Satellites

Energy source for the geodynamo

Heat flow across the core-mantle boundary

Ideal Solution theory

Inner core: composition

Interiors of planets and satellites

Magnetic Field of Mars

Melting temperature of iron in the core - theory

Radioactive isotopes, their decay in core & mantle

**Figures**

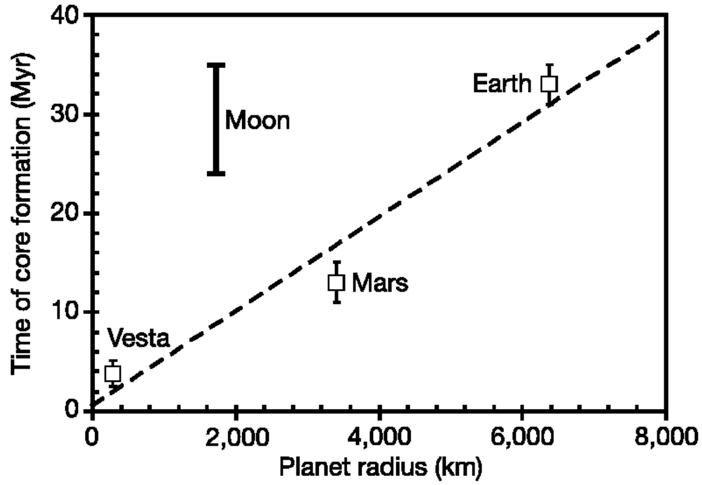


Figure 1: Model single-stage core formation age after solar system formation compared with present-day radius of body. From Kleine et al. (2002).

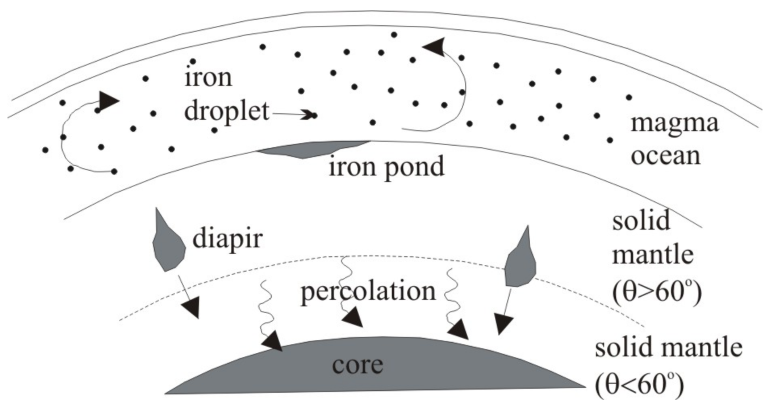


Figure 2: Mechanisms of core differentiation, after Stevenson (1990). $\theta$ is the dihedral angle.