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Constraints from material properties on the dynamics and evolution of Earth's core

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The Earth's magnetic field is powered from energy supplied by slow cooling and freezing of the liquid iron core. Core thermal history calculations have been hindered in the past by poor knowledge of the properties of iron alloys at the extreme pressures and temperatures pertaining in the core. This obstacle is now being overcome by developments in high pressure experiments and computational mineral physics. Here we review the relevant properties of iron alloys at core conditions and discuss their uncertainty and geophysical implications. Powerful constraints on core evolution are now possible, due partly to recent factor 2–3 upward revisions of the all-important electrical and thermal conductivities. This has dramatic implications for the thermal history of the entire Earth, not just the core: the inner core is very young, the core is cooling quickly, and was so hot in the past that the lowermost mantle

was almost certainly molten.

Turbulent motions in Earth's liquid outer core, a mixture of iron alloyed with lighter elements, generate the geomagnetic field through a dynamo process that converts kinetic energy into magnetic energy. Paleomagnetic observations show that the field has persisted for at least the last 3.5 billion years[1], which raises a fundamental question: how was the dynamo powered over this period? The standard model asserts that mantle convection cools the core by extracting heat across the core-mantle boundary (CMB); the resulting buoyancy forces drive vigorous convection that keeps the light element concentration almost uniform and the temperature close to adiabatic. Cooling leads to freezing of the liquid from the bottom up[2] because the melting curve $T_{\mathrm{m}}(P)$ increases more rapidly with pressure P than the adiabat $T_{\mathrm{a}}(P)$. As the solid inner core grows, latent heat is released and the light elements partition selectively into the outer core, reducing its density compared to pure iron[3] and providing a source of gravitational power[4]. Additional heating comes from the presence of any radiogenic elements.

In general, higher CMB heat flows lead to faster rates of cooling and inner core growth and provide more power for driving the dynamo (see Methods for mathematical details). Increasing the conductive heat loss Q_{a}, either through a larger thermal conductivity or temperature gradient, reduces the available power. Since all of the gravitational energy goes into generating magnetic field it makes the biggest contribution to determining the available dynamo power[5]. As well as the cooling rate, gravitational energy depends on the nature and mass concentration c of light elements and $\tau = \mathrm{d}T_{\mathrm{m}}/\mathrm{d}P - \partial T_{\mathrm{a}}/\partial P$, the difference between adiabatic and melting temperature gradients at the inner core boundary (ICB). Increasing c enhances the compositional density anomalies while reducing τ means that more inner core material freezes in unit time; for a given cooling rate both effects act to increase the gravitational energy.

Early models of core evolution used ideal solution theory to obtain c directly from density without needing to specify the species and represented τ in terms of one or more free parameters[6,7]. The numbers allowed an ancient inner core; the associated gravitational energy powered the geodynamo over most of Earth's history, negating any concerns over sustaining a dynamo powered by thermal convection alone. This scenario became untenable following an upward revision of T_a, which increased the adiabatic gradient and hence the heat Q_a conducted down the adiabat [see equations (1) and (2) below]. The prevailing view was that the inner core must be a young feature of the planet, around 1 billion years old[8], and that thermal convection alone could power the dynamo prior to inner core formation[9]. However, thermal history models still produced a wide range of results, owing to different choices for material properties rather than theoretical formulations[9].

The technical challenge of estimating core properties arises from the extreme pressures ($135 - 363$ GPa) and temperatures ($\sim$5000 K). This challenge is now being met by *ab initio* calculations and by diamond anvil cell and shock wave experiments, where available. *Ab initio* calculations deliver all the geophysically relevant parameters at the full range of core (P, c, T) conditions; they are ground truthed from experiments, which are usually conducted in more restrictive (P, c, T) regimes. Diamond anvil cell experiments are normally only available up to upper core (P, T) conditions, while shock wave experiments follow an equation of state defined by the physical properties of the material (the Hugoniot) and are therefore not able to explore the full (P, T) space relevant to the core (pre-heating or pre-compressing allows some movement in (P, T) space, but not enough to cover all the relevant conditions). Examples of validations of *ab initio* calculations on pure iron include the equation of state of the hexagonal close-packed crystal up to core pressures, both at room temperature[10–14] and on the Hugoniot[15,16], the speed of sound of the liquid[16,17], the isentropic compressibility and thermal expansivity of the solid on the Hugoniot[15,16], the phonon dispersions (vibrational frequencies of waves in crystals as function of wave-vector) of the body centered cubic crystal at ambient conditions[14,18], the density of states of hexagonal close-packed iron up to 150 GPa[19], the iron melting curve[17,20], and the ambient conditions electrical resistivity[21,22].

The most difficult quantities to calculate at core conditions happen to be the most critical for core and geodynamo models: thermal and electrical conductivities. Results have only been obtained recently[23–28], and turn out to be 2–3 times higher than conventional estimates[29,30] (called "low conductivities" henceforth) of $k = 28 - 46$ W m^{-1} K^{-1}. Crucially these new values ("high conductivities") have been obtained in both experiments and *ab initio* calculations. A very recent study[31] on a perfect iron crystal at ICB conditions suggests that a new effect (electron-electron scattering) would reduce the electrical conductivity back to old values that were estimated for the liquid[29]. The proposed importance of strong correlation effects appears at odds with previous work[32], so these results await both experimental and theoretical confirmation. Because of this we mainly focus on the high conductivity values, although the lower values are included for completeness.

Here we present a synthesis of core material properties. Parameter values are discussed, followed by their geophysical significance. A brief description of the *ab initio* methods is provided in Methods.

1 Material Properties for Earth's Core

The thermodynamic state of the core is determined by 3 intensive variables: Pressure P, mass concentration of species X, c_X, and temperature T. Pressure is very close to the enormous hydrostatic pressure, which is determined from seismology by integrating $\mathrm{d}P/\mathrm{d}r = -\rho g$ over radius r. Here ρ is density and g is gravity. Constraints on c_X and T are derived from the seismically-determined ICB density jump, $\Delta\rho$.

Part of the observed density jump[17], $\Delta\rho_{\mathrm{m}} = 0.24$ gm cc^{-1}, is due to the phase change at the ICB; the rest determines the excess concentration of light element in the outer core, which in turn affects the melting temperature and influences almost all terms in the energy and entropy budgets. Normal mode eigenfrequencies give a consistent result of $\Delta\rho = 0.8 \pm 0.2$ gm cc^{-1} but have low resolution[33] of about 400 km. Body waves have much better resolution of a few kilometers, but the estimates vary widely because PKiKP is a noisy phase[34–36]; they give an upper bound[36] of 1.1 gm cc^{-1}. There is also evidence for an anomalously dense layer in the lowermost 150 km of the outer core[37], which probably has a chemical origin[38]. Two explanations have been proposed: the layer could be a stable density-stratified zone of partial melt through which light elements pass by progressive melting and freezing[38], or parts of the inner core could be melting, releasing excess heavy liquid into the outer core[39]. In either case normal modes would measure the density difference between the inner core and main part of the outer core while body waves would measure the smaller difference between the solid inner core and the heavy liquid in the anomalous layer. We believe the normal mode estimates are more likely to represent the true compositional difference between the outer and inner cores. We consider the three values $\Delta\rho = 0.6$, 0.8 and 1.0 gm cc^{-1} spanning the range of published estimates. The 0.6 value corresponds to PREM[40].

Table 1 summarises our best estimates of core material properties for pure iron and the 3 values of $\Delta\rho$. Supplementary Table 1 is an extended version of Table 1 and Supplementary Tables 2–4 provide polynomial representations of depth-varying properties. Models are labeled by the corresponding core composition as described below. After composition we discuss thermal properties, followed by transport properties, which must be calculated for specific (P, c, T) conditions.

Composition is determined from the density (see Methods) and seismic velocities by comparing them with calculated values for mixtures of iron and candidate siderophile elements: Si and O because of their abundance and S because of its presence in iron meteorites, which are thought to be remnants of planetary cores. Other elements, e.g. H, have been proposed[41] but their properties in iron mixtures have not yet been explored extensively. The core also probably contains some Ni; however, recent experiments found that adding up to 10% of Ni does not change the hexagonal close-packed crystal structure of the solid[42], while *ab initio* calculations suggest that at high T the seismic properties of Fe-Ni alloys are almost indistinguishable from those of pure iron[43]. Recent studies of core composition[44–46] conclude that the light elements are likely to be Si, S, and O with negligible amounts of H and C. *Ab initio* calculations for Fe-S, Fe-Si and Fe-O mixtures show that S and Si partition almost equally between solid and liquid, while almost all the O goes into the liquid[14,45]. The behaviour of S and Si are very similar[14] so we use a Fe-Si-O mixture in this review. Mass concentrations of species X for the solid and liquid, c_X^s and c_X^l respectively, are given in section 1 of Table 1; each model is named after the corresponding molar concentration.

Temperature. Light element X depresses the melting temperature for pure iron, T_{m}, by an amount ΔT_{X}. Of particular importance are conditions near the ICB (radius $r = r_{\mathrm{i}}$, $P = 330$ GPa). The large volume of work on T_{m} is summarised elsewhere[20,47]. Some studies have shown encouraging agreement, with $T_{\mathrm{m}}(r_{\mathrm{i}}) = 6350 \pm 300$ K predicted by diamond anvil cell experiments[20,47] up to 82 GPa and 200 GPa respectively, shock experiments[48] at 225–260 GPa and *ab initio* calculations[14,49] at 330 GPa. This value is used in Section 2 of Table 1. Other calculations[50,51] have found $T_{\mathrm{m}}(r_{\mathrm{i}}) = 7100$ K and $T_{\mathrm{m}}(r_{\mathrm{i}}) = 5400$ K respectively, but these only used *ab initio* indirectly by fitting an interatomic potential which has different melting properties from those of the fully *ab initio* system[52].

Along with T_{m} and the core chemistry model, the entropy of melting for pure iron ΔS is needed to determine ΔT_{X} at the ICB[49]. The core temperature at the ICB, T_{i}, equals the melting temperature of the mixture; the values in section 2 of Table 1 are calculated from $T_{\mathrm{i}} = T_{\mathrm{m}} + \Delta T_{\mathrm{O}} + \Delta T_{\mathrm{Si}}$. The latent heat L released on freezing the inner core is $L = T_{\mathrm{m}} \Delta S$ (section 2 of Table 1).

In regions where convection is active the outer core temperature follows an adiabat, given by

$$T_{\rm a} = T_{\rm i} \exp\left(-\int_{r_{\rm i}}^{r} \frac{\rho g \gamma}{K_s} {\rm d}r\right), \tag{1}$$

where γ is the thermodynamic Grüneisen parameter. Note that $\partial T_{\rm a}/\partial r = -\rho g \gamma T_{\rm a}/K_s$. The bulk modulus, K_s, and gravity, g, are calculated directly in *ab initio* methods and are very similar to PREM. *Ab initio* calculations have found that $\gamma \approx 1.5$ at the CMB and remains constant[17] (to within the accuracy of the calculations) or decreases slightly[53,54] with depth. The depth variation reduces $\partial T_{\rm a}/\partial r$ and increases $\tau = {\rm d}T_{\rm m}/{\rm d}P - \partial T_{\rm a}/\partial P$, but makes little difference to $T_{\rm a}$. Depth variation of $T_{\rm a}$ is therefore well-constrained. The three adiabats used in the core evolution calculations below are shown in Figure 1; values for the CMB and ICB gradients are given in section 3 of Table 1. In the inner core, $T_{\rm a}$ was assumed to be close to isothermal[27].

The thermal and chemical expansion coefficients, $\alpha_T = \rho^{-1}(\partial \rho/\partial T)_{P,c}$ and $\alpha_c = -\rho^{-1}(\partial \rho/\partial c)_{P,T}$, determine the buoyancy forces arising from thermal and compositional anomalies. α_T can be obtained from a number of thermodynamic relations, e.g. $\alpha_T = \gamma \rho C_p / K_s$. *Ab initio* calculations have found the specific heat $C_p = 700 - 800$ J kg^{-1} K^{-1} independent of radius[54], in agreement with theory[55] and hence α_T is a decreasing function of depth[55,56] because of the factor ρ/K_s. The compositional expansion coefficient α_c is different for each element; values obtained at present ICB (P, T) conditions[49] are given in Table 1.

Transport Properties. The geophysical importance of core thermal (k) and electrical (σ) conductivities is discussed below. σ is easier to obtain and is sometimes used to infer k through the Wiedemann-Franz law, although there are situations when this relation does not hold (see Methods). Recent estimates of k and σ for pure iron[23,24] are 3–5 times higher at the CMB than previous estimates[29,30] and increase by a factor of 1.5 to the ICB. Mixtures have also been studied, though using different compositions and adiabats. Despite this, and the different methods used, the studies all find k at the CMB in the range 80–110 W m^{-1} K^{-1}, increasing up to 140–160 W m^{-1} K^{-1} at the ICB[23,25,26] (Figure 1). There is a jump in both k and σ at the ICB and a small increase across the inner core[27].

Mass diffusion coefficients D_X relate the concentration gradient of species X to the diffusive flux of that species. Recent estimates[25,57] of D_O and D_{Si} agree with previous calculations at CMB pressures[58] and show a factor 1.5 increase to the ICB. In core evolution models D_X enters the barodiffusion term, which describes the entropy generated by diffusion of light elements down the ambient pressure gradient. The effect is measured by the barodiffusive coefficients α_X^D, which are calculated using the values of D_X and $(\partial\mu/\partial c_X)_{P,T}$ in Table 1, where μ is the chemical potential[58]. Barodiffusion is small enough to be neglected in the entropy budget[9,58,59], but might play a dynamical role near the top of the core (see the "stratification" subsection below).

The kinematic viscosity ν plays a key role in the dynamics of rotating fluids[60], but is less important for determining long-term core evolution. Recent *ab initio* estimates[25,57] of ν are given in Table 1 for the present core chemistry model; they are in line with older values[61].

2 Geophysical Implications of Revised Core Properties

Core Energy Budget. The dynamo entropy E_{J} represents the work done by buoyancy forces that goes into generating magnetic field[5] and is therefore crucial for assessing the viability of dynamo action. Both E_{J} and the CMB heat flow Q_{cmb} are related to the core cooling rate through the material properties described above: higher heat flow yields faster cooling and higher E_{J} (see Methods for details). The cooling rate determines the inner core age. Mantle convection sets the CMB heat flow and various lines of evidence suggest $Q_{\mathrm{cmb}} = 5 - 15$ TW at present[62,63]. E_{J} could be calculated directly if we had detailed knowledge of the magnetic field throughout the core; however, the main field contributions to E_{J} occur at scales that cannot be observed[64] and so E_{J} is determined from Q_{cmb} for the present-day. On longer timescales, where both Q_{cmb} and E_{J} are hard to estimate, the constraint $E_{\mathrm{J}} \geq 0$ can be used to calculate lower bounds on the cooling rate. All parameters values are given in Table 1; the most important are $\Delta\rho$ and k as we will show.

Increasing $\Delta\rho$ increases the outer core light element concentration and reduces the adiabatic gradient (because $\partial T_{\mathrm{a}}/\partial r$ is proportional to T_{a}), allowing the same E_{J} to be balanced with a lower cooling rate and hence lower Q_{cmb} (Figure 2). For a plausible value[65] of $E_{\mathrm{J}} = 400$ MW K^{-1}, increasing $\Delta\rho$ from 0.6 to 1.0 gm cc^{-1} reduces the required CMB heat flow by ≈ 2 TW with low k and ≈ 4 TW with high k.

Increasing k increases the amount of heat conducted away down the adiabatic gradient, and hence reduces the dynamo efficiency (Figure 2). The stability of core convection also depends critically on k. The total adiabatic heat flow is

$$Q_{\mathrm{a}} = 4\pi r_{\mathrm{o}}^2 k(r_{\mathrm{o}}) \left. \frac{\partial T_{\mathrm{a}}}{\partial r} \right|_{r=r_{\mathrm{o}}}, \tag{2}$$

When $Q_{\rm cmb} > Q_{\rm a}$ the whole core is superadiabatic and thermal convection occurs everywhere; when $Q_{\rm cmb} < Q_{\rm a}$ the top of the core is subadiabatic and stable to thermal convection. For a low value of $k = 28$ W m^{-1} K^{-1} the core is thermally unstable ($Q_{\rm cmb} > Q_{\rm a}$) and can generate a magnetic field ($E_{\rm J} \geq 0$) for all estimates of present-day CMB heat flow (Figure 2). For the high values of k dynamo action requires a minimum of 5.5–7.5 TW, while the top of the core is likely to be thermally stable unless $Q_{\rm cmb} \approx 15$ TW. This is very high, around one third of the total heat leaving Earth's surface[66]. Maintaining $E_{\rm J} = 400$ MW K^{-1} with the high k values requires $Q_{\rm cmb} =$ 9–13 TW with composition driving convection against thermal stratification in the uppermost core (Figure 2).

Thermal History. To evidence the effect of material properties on predictions of past core evolution we set $E_{\rm J} = 0$ prior to inner core formation and specify $Q_{\rm cmb}$ during inner core growth. This prescription[9,59,63] ensures that $Q_{\rm cmb} > Q_{\rm a}$, consistent with the modeling assumptions (see Methods), and produces conservative estimates of the cooling rate, core temperature and inner core age. Figure 3 shows predicted inner core age and CMB temperature ($T^{3.5{\rm Ga}}$) and CMB heat flow ($Q^{3.5{\rm Ga}}$) at 3.5 Ga, the time of the earliest paleomagnetic measurement[1]. The influence of radiogenic heating is demonstrated by adding 300 ppm of ^{40}K at the present day, which likely represents an extreme scenario[44,63]. The shaded temperature range of 4150 ± 150 K corresponds to present estimates of the lower mantle solidus temperature[67]; core temperatures exceeding this range suggest partial melting in past.

Low k models predict inner core ages of $\sim$1 Ga or more, CMB heat flows below 10 TW over the last 3.5 Ga and ancient core temperatures at or above the lower mantle solidus estimates. With the high k values there is little doubt that the lowermost mantle would have been partially molten in the past. Moreover, the high k models consistently yield inner core ages of $\sim$0.6 Ga or younger. Radiogenic heating does little to change the results. Figure 3 also shows favoured models from four recent studies[63,68–70] that use the high k values and impose different constraints on the time-variation of E_{J}. A consistent picture emerges in which 1) the inner core is at most 500–600 million years old; 2) ancient core temperatures greatly exceeded present estimates of the lower mantle solidus; 3) high ancient CMB heat flows were needed to power the early geodynamo.

Increasing $\Delta\rho$ from 0.6 gm cc^{-1} to 1.0 gm cc^{-1} can produce a 400–600 K decrease in $T^{3.5\mathrm{Ga}}$ and a 200–400 Myr increase in the inner core age, depending on the details on the model (Figure 3). Figure 4 shows how the results from a single reference case in Figure 3 are influenced by individually varying values for several material properties compared to the numbers in Table 1. Where errors are not reported a $\pm 10\%$ variation is assumed, which is likely to be larger than errors in the *ab initio* calculations[17,56]. Individually changing α_c or L by $\pm 10\%$, C_p to the values of a previous study[55], core density from PREM to AK135[71], or the melting curve to a recent experimental profile[20] (denoted T_{m}^{A}) each make little difference. Using a depth-variable γ (denoted γ^{I})[54] makes a small change to the inner core age but barely changes $T^{3.5\mathrm{Ga}}$. The biggest changes arise from varying k and allowing for the ± 300 K uncertainty in T_{i}. Combining the variations to give the youngest (oldest) inner core yields changes of +(-) 400 K in $T^{3.5\mathrm{Ga}}$ and -(+) 150 Myrs in inner core age compared to the reference model, which is a comparable effect to uncertainty in $\Delta\rho$ alone.

Stratification Beneath the CMB. Observed variations in the magnetic field only reflect changes at the top of the core and so the dynamic stability of this region is an important issue. Stratified layers are dynamically very different from convecting regions: they suppress radial motion and support a different suite of waves[72]. In the absence of chemical or boundary effects, subadiabatic conditions at the top of the core (Figure 2) should result in stable stratification. Compositional convection could overcome this stratification and mix the excess heat downwards, restoring adiabatic conditions everywhere[73]. Alternatively, light elements could enhance thermal stratification if they are emplaced at the top of the core early in Earth's history[74] or pool beneath the CMB over time. Pooling could arise from light element transfer across the CMB[75], by barodiffusion of light elements up the ambient pressure gradient[76], or by the transfer of chemically distinct blobs from the ICB[74,77].

Pooling mechanisms produce layers of $\sim$100 km depth[75,76], comparable to values inferred from geomagnetism[78], but thinner than recent seismic estimates[79]. Whether compositional convection can overcome thermal stratification requires detailed analyses of the different buoyancy sources[26,80,81]. Two recent studies[78,81] find a thermochemically stable layer of $\sim$100 km for a CMB heat flow of $\approx$13 TW, compatible with current $Q_{\rm cmb}$ estimates[62]. Estimates of the associated density gradients from the recently-proposed thermal/chemical stable layers yield Brunt frequencies of $O(1)$ day[75,76,81], eliminating any longer-period vertical motion.

Density anomalies associated with core motions are so small that convection is unlikely to entrain or penetrate a stable layer[26,72,75,76]. The effect on a stable layer of thermal anomalies in the lowermost mantle is not so clear. The large-scale pattern of CMB heat flow can be constructed by assuming that observed seismic velocity variations represent thermal heterogeneity. The strength of the lateral variations is measured by the parameter $q^* = (q^{\mathrm{max}} - q^{\mathrm{min}})/(q_{\mathrm{cmb}} - q_{\mathrm{a}})$, the ratio of peak-to-peak boundary heat flow variations to the mean superadiabatic heat flow per unit area. Mantle convection simulations[82] have estimated $q^* \approx 2$, but did not appear to subtract the adiabat. The high values of k increase q_{a} and hence q^*, further strengthening the effect.

Geodynamo simulations with $q^* \approx 1$ produce flows with persistent downwellings below regions of high CMB heat flow that concentrate magnetic flux there, producing field morphologies that are similar to the historical geomagnetic field[83,84]. These effects will be amplified when convection is weak at the top. Boundary-driven radial motions may generate flow in a stratified layer[81], as has been observed in non-magnetic simulations with weak stratification[85]. Geodynamo simulations that combine strong stratification and strong boundary forcing ($q^* \gg 1$) are needed to establish which dynamics win out.

The depth increase of k opens up the possibility that the very top of the core is superadiabatic, with a stable layer directly beneath[26,80]. The conditions required to form such a layer are sensitive to the $T_{\mathrm{a}}(r)$ and $k(r)$ profiles; the models in this review do not produce such an effect.

Magnetic timescales. Revised core viscosity and diffusivities (Table 1) are too small to be used in present geodynamo simulations. This situation is unlikely to change in the next ten years[60]. However, changes to the electrical conductivity σ are significant. The new (high) values of σ give a magnetic diffusivity of $\eta = 0.7$ m^2 s^{-1} at the CMB and $\eta = 0.6$ m^2 s^{-1} at the ICB compared to $\eta = 1.6$ m^2 s^{-1} using a low value[29] of $\sigma = 5 \times 10^5$ S m^{-1}. Lowering η raises the Magnetic Reynold's number $Rm = Ur_\mathrm{o}/\eta$ from ≈700 to ≈1500, where U is the root mean square velocity at the top of the core[25,26]. Rm must be sufficiently large to generate a magnetic field by dynamo action. Decreasing η makes dynamo action possible with slower flows.

The time for a dipole magnetic field (the slowest decaying mode) to decay in a uniform sphere of radius r_o, the dipole decay time $\tau_d = r_\mathrm{o}^2/\pi^2\eta$, is increased from 25 kyrs to 55 kyrs with the revised σ values. This result changes interpretations of all geomagnetic observations in terms of diffusion processes. In particular, polarity reversals of the field, which take 1–10 kyrs to complete, now appear fast on the diffusion timescale: $\tau_d = 10$ kyr for the inner core, comparable to the timescale of reversal transition. Whether this is coincidence or a characteristic that distinguishes reversals from excursions[86] (where the new polarity is not retained) remains to be tested with modern geodynamo models.

Inner Core Convection. Seismic observations have revealed surprising structural complexity in the inner core, including hemispherical and radial variations in velocity and anisotropy[37]. Much recent work has focused on explaining these observations by solid-state convection[87]. Thermal convection requires the inner core to be superadiabatic; with the high values of $k \sim 200$ W m^{-1} K^{-1} (Figure 1) this requires $Q_\mathrm{cmb} = 30 - 60$ TW at the present-day[27,70,88], at least two thirds of the surface heat flow[66]. Just after inner core nucleation, 500–600 Myr ago (Figure 3), an estimated 30 TW is needed[27]. Mantle heat sources are unlikely to have changed significantly in this period[89]; 30 TW probably represents at least half of Earth's total heat budget at this time.

Inner core convection could be driven compositionally if less light element partitions into it over time. Compositionally unstable conditions may have arisen once the inner core grew beyond $O(10)$ km, but probably have not persisted to the present day[59,70]. The case of thermochemical buoyancy is complicated by possible double-diffusive effects; initial studies indicate that the net buoyancy force is stabilising[90]. Overall it seems that inner core convection, either in the plume[87] or translation[39,91] regimes, is unlikely at present. This is consistent with a recent review that favours texturing mechanisms arising from magnetic coupling or heterogeneous growth due to enhanced equatorial heat loss[88]. If heterogeneous ICB heat flow is related to recent geomagnetic phenomena such as weak secular variation in the Pacific hemisphere[92] or long-term tilt of the dipole axis[93] then another mechanism (aside from convection) may be needed to explain the origin of the forcing.

3 Core Dynamics and Evolution with High Conductivities

The material properties of liquid iron alloys at high pressures and temperatures are now sufficiently well-known to draw robust conclusions about the long-term evolution of the core. Recent calculations with the new (high) conductivities find that 1) The inner core age is less than 500-600 Ma[24,59,63,80]; 2) high early CMB heat flow and corresponding core temperatures that significantly exceeded present estimates of the lower mantle solidus temperature[59,63,68,94] imply partial melting of the lowermost mantle in the past; 3) the present-day core is subadiabatic at the top and may be stably stratified[24,26,80]. Prior to the high conductivity estimates, models predicted inner core nucleation 1 billion years ago[8], early core temperatures comparable to the lower mantle solidus[9], and superadiabatic conditions throughout the core at the present-day.

In terms of geophysical significance the most uncertain properties are the iron melting curve T_m and the ICB density jump $\Delta\rho$. However, the preceding conclusions will hold unless $\Delta\rho$ or T_m have been drastically underestimated. Core composition is also important: we have used an Fe-Si-O model, but other species such as H and C have been proposed. The effects of other putative light elements can now be investigated routinely using *ab initio* methods and the results evaluated against geophysical constraints. The viability of a given composition is no longer a matter for speculation. Finally, there is still some debate over the conductivity. The implications of old (low) conductivity values are shown in Figures 2 and 3. We favour the high values and discuss their implications below.

Revised core evolution models indicate that powering the dynamo around 3.5 Ga required a minimum $Q_\mathrm{cmb} \approx 15 - 25$ TW to be extracted from the core by a partially molten lower mantle. The actual required Q_cmb at this time was likely much greater. Internal heat production within a magma ocean due to latent heat release and/or radiogenic sources will insulate the core, further exacerbating the heat problem[95]. It has been proposed that the insulating effect was so drastic as to delay the onset of the core dynamo until $\sim$2 Ga, with the magma ocean generating the field before this time[96]. Whether cooling alone is sufficient to power the early dynamo is currently an open question; indeed, the search for alternative energy sources has already begun[97].

At present the uppermost core is subadiabatic unless Q_cmb has been underestimated; however, this seems unlikely based on the power requirements for mantle convection[62]. The magnetic field is generated by vigorous convection deep within the core, powered by latent heat release and gravitational energy. If light elements pool at the CMB the top of the core will be stably stratified. Lateral variations in CMB heat flow are superimposed on the stratified layer. Geomagnetic data are presently unable to unambiguously identify a stable layer[98,99], although a recent constraint on core electrical conductivity from long-term dipole field variations is consistent with the high conductivity estimates that argue in favour of stratification[100]. In isolation, both a stable layer and lateral heat flow variations can explain prominent features of the present geomagnetic field: wave motions in a $\sim$100 km-thick stable layer can account for short-period fluctuations in the dipole field[78]; regions of high CMB heat flow can concentrate magnetic field lines, producing the four dominant high-latitude flux patches[83]; low heat flow beneath the Pacific can explain the weak secular variation there[85]. Progress towards a coherent dynamical model of the present-day core requires 1) a coherent seismological picture of core stratification; 2) explaining recent geomagnetic secular variation in terms of stable layer dynamics and; 3) analysis of the interaction between a stable region and CMB heat flow variations. The origin of a stable layer poses yet more fascinating challenges for future research.

1. Tarduno, J. *et al.* Geodynamo, solar wind, and magnetopause 3.4 to 3.45 billion years ago. *Science* **327**, 1238–1240 (2010).
2. Jacobs, J. The Earth's inner core. *Nature* **172**, 297–300 (1953).
3. Birch, F. Elasticity and the constitution of Earth's interior. *J. Geophys. Res.* **66**, 227–286 (1952).
4. Braginsky, S. Structure of the F layer and reasons for convection in the Earth's core. *Sov.*

Phys. Dokl. **149**, 8–10 (1963).

5. Lister, J. & Buffett, B. The strength and efficiency of thermal and compositional convection in the geodynamo. *Phys. Earth Planet. Int.* **91**, 17–30 (1995).
6. Stevenson, D., Spohn, T. & Schubert, G. Magnetism and thermal evolution of the terrestrial planets. *Icarus* **54**, 466–489 (1983).
7. Mollett, S. Thermal and magnetic constraints on the cooling of the Earth. *Geophys. J. R. Astr. Soc.* **76**, 653–666 (1984).
8. Labrosse, S., Poirier, J.-P. & Le Moeul, J.-L. The age of the inner core. *Earth Planet. Sci. Lett.* **190**, 111–123 (2001).
9. Nimmo, F. Thermal and compositional evolution of the core. In Schubert, G. (ed.) *Treatise on Geophysics, Vol. 9*, 217–241 (Elsevier, Amsterdam, 2007).
10. Söderlind, P., Moriarty, J. A. & Wills, J. M. First-principles theory of iron up to Earth-core pressures: Structural, vibrational, and elastic properties. *Phys. Rev. B* **53**, 14063 (1996).
11. Vočadlo, L., de Wijs, G. A., Kresse, G., Gillan, M. & Price, G. D. First principles calculations on crystalline and liquid iron at Earth's core conditions. *Faraday Discuss.* **106**, 205 (1997).
12. Mao, K., Wu, Y., Chen, L. C., Shu, J. F. & Jephcoat, A. P. Static compression of iron to 300 GPa and $\mathrm{Fe}_{0.8}\mathrm{Ni}_{0.2}$ alloy to 260 GPa: implications for the composition of the core. *J. Geophys. Res.* **95**, 21737 (1990).

13. Dewaele, A. *et al.* Quasihydrostatic equation of state of iron above 2 Mbar. *Phys. Rev. Lett.* **97**, 215504 (2006).

14. Alfè, D., Gillan, M. J. & Price, G. D. Temperature and composition of the Earth's core. *Contemp. Phys.* **48**, 63–80 (2007).

15. Alfè, D., Price, G. & Gillan, M. Thermodynamics of hexagonal close packed iron under Earth's core conditions. *Phys. Rev. B* **64**, 045123 (2001).

16. Brown, J. M. & McQueen, R. G. Phase transitions, Grüneisen parameter, and elasticity for shocked iron between 77 GPa and 400 GPa. *J. Geophys. Res.* **91**, 7485 (1986).

17. Alfè, D., Price, G. & Gillan, M. Iron under Earth's core conditions: Liquid-state thermodynamics and high-pressure melting curve from *ab initio* calculations. *Phys. Rev. B* **65**, 165118 (2002).

18. Brockhouse, B. N., Abou-Helal, H. E. & Hallman, E. D. Lattice vibrations in iron at 296 K. *Solid State Commun.* **5**, 211 (1967).

19. Mao, H. K. *et al.* Phonon density of states of iron up to 153 GPa. *Science* **292**, 914 (2001).

20. Anzellini, S., Dewaele, A., Mezouar, M., Loubeyre, P. & Morard, G. Melting of iron at Earth's inner core boundary based on fast x-ray diffraction. *Science* **340**, 464–466 (2013).

21. Alfè, D., Pozzo, M. & Desjarlais, M. Lattice electrical resistivity of magnetic bcc iron from first-principles calculations. *Phys. Rev. B* **85**, 1–5 (2012).

22. Weiss, R. J. & Marotta, A. S. Spin-dependence of the resistivity of magnetic metals. *J. Phys. Chem. Solids* **9**, 302–308 (1959).

23. de Koker, N., Steinle-Neumann, G. & Vojtech, V. Electrical resistivity and thermal conductivity of liquid Fe alloys at high P and T and heat flux in Earth's core. *Proc. Natl. Acad. Sci.* **109**, 4070–4073 (2012).

24. Pozzo, M., Davies, C., Gubbins, D. & Alfè, D. Thermal and electrical conductivity of iron at Earth's core conditions. *Nature* **485**, 355–358 (2012).

25. Pozzo, M., Davies, C., Gubbins, D. & Alfè, D. Transport properties for liquid silicon-oxygen-iron mixtures at Earth's core conditions. *Phys. Rev. B* **87**, 014110 (2013).

26. Gomi, H. *et al.* The high conductivity of iron and thermal evolution of the Earth's core. *Phys. Earth Planet. Int.* **224**, 88–103 (2013).

27. Pozzo, M., Davies, C., Gubbins, D. & Alfè, D. Thermal and electrical conductivity of solid iron and iron-silicon mixtures at Earth's core conditions. *Earth Planet. Sci. Lett.* **393**, 159–164 (2014).

28. Ohta, K. e. a. Measurements of electrical and thermal conductivity of iron under Earth's core conditions **Abstract AGU Fall Meeting 2014** (2014).

29. Stacey, F. & Anderson, O. Electrical and thermal conductivities of Fe-Ni-Si alloy under core conditions. *Phys. Earth Planet. Int.* **124**, 153–162 (2001).

30. Stacey, F. & Loper, D. A revised estimate of the conductivity of iron alloy at high pressure and implications for the core energy balance. *Phys. Earth Planet. Int.* **161**, 13–18 (2007).

31. Zhang, P., Cohen, R. & Haule, K. Effects of electron correlations on transport properties of iron at Earth's core conditions. *Nature* **517**, 605–607 (2015).

32. Pourovskii, L., Mravlje, J., Ferrero, M., Parcollet, O. & Abrikosov, I. Impact of electronic correlations on the equation of state and transport in ϵ-Fe. *Phys. Rev. B* **90**, 155120 (2014).

33. Masters, G. & Gubbins, D. On the resolution of density within the Earth. *Phys. Earth Planet. Int.* **140**, 159–167 (2003).

34. Cao, A. M. & Romanowicz, B. Constraints on density and shear velocity contrasts at the inner core boundary. *Geophys. J. Int.* **157**, 1146–1151 (2004).

35. Koper, K. D. & Dombrovskaya, M. Seismic properties of the inner core boundary from PKiKP/P amplitude ratios. *Earth Planet. Sci. Lett.* **237**, 680–694 (2005).

36. Tkalčić, H., Kennett, B. & Cormier, V. On the inner–outer core density contrast from PKiKP/PcP amplitude ratios and uncertainties caused by seismic noise. *Geophys. J. Int.* **179**, 425–443 (2009).

37. Souriau, A. Deep Earth structure – the Earth's cores. In Schubert, G., Romanowicz, B. & Dziewonski, A. (eds.) *Treatise on geophysics*, vol. 1, chap. 19, 655–693 (Elsevier, Amsterdam, 2007).

38. Gubbins, D., Masters, G. & Nimmo, F. A thermochemical boundary layer at the base of Earth's outer core and independent estimate of core heat flux. *Geophys. J. Int.* **174**, 1007–1018 (2008).

39. Alboussière, T., Deguen, R. & Melzani, M. Melting-induced stratification above the Earth's inner core due to convective translation. *Nature* **466**, 744–747 (2010).

40. Dziewonski, A. & Anderson, D. Preliminary Reference Earth Model. *Phys. Earth Planet. Int.* **25**, 297–356 (1981).

41. Nomura, R. *et al.* Low core-mantle boundary temperature inferred from the solidus of pyrolite. *Science* **343**, 522–524 (2014).

42. Tateno, S., Hirose, K., Komabayashi, T., Ozawa, H. & Ohishi, Y. The structure of Fe-Ni alloy in Earth's inner core. *Geophys. Res. Lett.* **39**, L12305 (2012).

43. Martorell, B., Brodholt, J., Wood, I. G. & Vočadlo, L. The effect of nickel on the properties of iron at the conditions of the Earth's inner core: *Ab initio* calculations of seismic wave velocities of Fe-Ni alloys. *Earth Planet. Sci. Lett.* **365**, 143–151 (2013).

44. Hirose, K., Labrosse, S. & Hernlund, J. Compositional state of Earth's core. *Annual Review of Earth and Planetary Sciences* **41**, 657–691 (2013).

45. Morard, G., Andrault, D., Antonangeli, D. & Bouchet, J. Properties of iron alloys under the Earth's core conditions. *C. R. Geoscience* **346**, 130–139 (2014).

46. Badro, J., Côté, A. & Brodholt, J. A seismologically consistent compositional model of Earth's core. *Proc. Natl. Acad. Sci.* **111**, 7542–7545 (2014).

47. Jackson, J. *et al.* Melting of compressed iron by monitoring atomic dynamics. *Earth Planet. Sci. Lett.* **362**, 143–150 (2013).

48. Nguyen, J. & Holmes, N. Melting of iron at the physical conditions of the Earth's core. *Nature* **427**, 339–342 (2004).

49. Alfè, D., Gillan, M. & Price, G. *Ab initio* chemical potentials of solid and liquid solutions and the chemistry of the Earth's core. *J. Chem. Phys.* **116**, 7127–7136 (2002).

50. Belonoshko, A., Ahuja, R. & Johansson, B. Quasi–*ab Initio* molecular dynamic study of Fe melting. *Phys. Rev. Lett.* **84**, 3638–3641 (2000).

51. Laio, A., Bernard, S., Chiarotti, G. & Scandolo, E., S. Tosatti. Physics of iron at Earth's core conditions. *Science* **287**, 1027–1030 (2000).

52. Alfè, D., Gillan, M. & Price, G. Complementary approaches to the *ab initio* calculation of melting properties. *J. Chem. Phys.* **116**, 6170–6177 (2002).

53. Stacey, F. Thermodynamics of the Earth. *Rep. Prog. Phys.* **73**, 1–23 (2010).

54. Ichikawa, H., Tsuchiya, T. & Tange, Y. The P-V-T equation of state and thermodynamic properties of liquid iron. *J. Geophys. Res.* **119**, 240–252 (2014).

55. Stacey, F. Core properties, physical. In Gubbins, D. & Herrero-Bervera, E. (eds.) *Encyclopedia of Geomagnetism and Paleomagnetism*, 91–94 (Springer, 2007).

56. Gubbins, D., Alfe, D., Masters, G., Price, G. & Gillan, M. Can the Earth's dynamo run on heat alone? *Geophys. J. Int.* **155**, 609–622 (2003).

57. Ichikawa, H. & Tsuchiya, T. Atomic transport property of Fe-O liquid alloys in the Earth's outer core P,T condition. *Phys. Earth Planet. Int.* dx.doi.org/10.1016/j.pepi.2015.03.006 (2015).

58. Gubbins, D., Alfè, D., Masters, G., Price, G. & Gillan, M. Gross thermodynamics of two-component core convection. *Geophys. J. Int.* **157**, 1407–1414 (2004).

59. Davies, C. Cooling history of Earth's core with high thermal conductivity. *Phys. Earth Planet. Int.* doi:10.1016/j.pepi.2015.03.007 (2015).

60. Roberts, P. & King, E. On the genesis of the Earth's magnetism. *Rep. Prog. Phys.* **76**, 1–55 (2013).

61. de Wijs, G. *et al.* The viscosity of liquid iron at the physical conditions of the Earth's core. *Nature* **392**, 805–807 (1998).

62. Lay, T., Hernlund, J. & Buffett, B. Core-mantle boundary heat flow. *Nat. Geosci.* **1**, 25–32 (2009).

63. Nimmo, F. Energetics of the core. In Schubert, G. (ed.) *Treatise on Geophysics 2nd Edn, Vol. 9*, 31–65 (Elsevier, Amsterdam, 2014).

64. Jackson, A. & Livermore, P. On Ohmic heating in the Earth's core I: nutation constraints. *Geophys. J. Int.* **177**, 367–382 (2009).

65. Roberts, P., Jones, C. & Calderwood, A. Energy fluxes and ohmic dissipation in the Earth's core. In *Earth's core and lower mantle*, 100–129 (Contributions from the SEDI 2000, The 7th Symposium, 2003).

66. Davies, J. & Davies, D. Earth's surface heat flux. *Solid Earth* **1**, 5–24 (2010).

67. Andrault, D. *et al.* Solidus and liquidus profiles of chondritic mantle: Implication for melting of the Earth across its history. *Earth Planet. Sci. Lett.* **304**, 251–259 (2011).

68. Nakagawa, T. & Tackley, P. Influence of combined primordial layering and recycled MORB on the coupled thermal evolution of Earth's mantle and core. *Geochem. Geophys. Geosys.* **15**, 619–633 (2014).

69. Driscoll, P. & Bercovici, D. On the thermal and magnetic histories of Earth and Venus: Influences of melting, radioactivity, and conductivity. *Phys. Earth Planet. Int.* **236**, 36–51 (2014).

70. Labrosse, S. Thermal and compositional stratification of the inner core. *C. R. Geosci.* **346**, 119–129 (2014).

71. Kennett, B., Engdahl, E. & Buland, R. Constraints on seismic velocities in the Earth from traveltimes. *Geophys. J. Int.* **122**, 108–124 (1995).

72. Braginsky, S. Dynamics of the stably stratified ocean at the top of the core. *Phys. Earth Planet. Int.* **111**, 21–34 (1999).

73. Loper, D. Some thermal consequences of a gravitationally powered dynamo. *J. Geophys. Res.* **831**, 5961–5970 (1978).

74. Helffrich, G. & Kaneshima, S. Causes and consequences of outer core stratification. *Phys. Earth Planet. Int.* **223**, 2–7 (2013).

75. Buffett, B. & Seagle, C. Stratification of the top of the core due to chemical interactions with the mantle. *J. Geophys. Res.* **115**, B04407 (2010).

76. Gubbins, D. & Davies, C. The stratified layer at the core-mantle boundary caused by barodiffusion of Oxygen, Sulphur and Silicon. *Phys. Earth Planet. Int.* **215**, 21–28 (2013).

77. Moffatt, H. & Loper, D. The magnetostrophic rise of a buoyant parcel in the Earth's core. *Geophys. J. Int.* **117**, 394–402 (1994).

78. Buffett, B. Geomagnetic fluctuations reveal stable stratification at the top of the Earth's core. *Nature* **507**, 484–487 (2014).

79. Helffrich, G. & Kaneshima, S. Outer-core compositional stratification from observed core wave speed profiles. *Nature* **468**, 807–809 (2010).

80. Labrosse, S. Thermal evolution of the core with a high thermal conductivity. *Phys. Earth Planet. Int.* doi:10.1016/j.pepi.2015.02.002 (2015).

81. Gubbins, D., Alfè, D., Davies, C. & Pozzo, M. On core convection and the geodynamo: Effects of high electrical and thermal conductivity. *Phys. Earth Planet. Int.* doi:10.1016/j.pepi.2015.04.002 (2015).

82. Nakagawa, T. & Tackley, P. Lateral variations in CMB heat flux and deep mantle seismic velocity caused by a thermal-chemical-phase boundary layer in 3D spherical convection. *Earth Planet. Sci. Lett.* **271**, 348–358 (2007).

83. Olson, P. & Christensen, U. The time-averaged magnetic field in numerical dynamos with non-uniform boundary heat flow. *Geophys. J. Int.* **151**, 809–823 (2002).

84. Aubert, J., Amit, H., Hulot, G. & Olson, P. Thermochemical flows couple the Earth's inner core growth to mantle heterogeneity,. *Nature* **454**, 758–761 (2008).

85. Gibbons, S. & Gubbins, D. Convection in the Earth's core driven by lateral variations in core-mantle boundary heat flux. *Geophys. J. Int.* **142**, 631–642 (2000).

86. Hollerbach, R. & Jones, C. Influence of the Earth's inner core on geomagnetic fluctuations and reversals. *Nature* **365**, 541–543 (1993).

87. Deguen, R. Structure and dynamics of Earth's inner core. *Earth Planet. Sci. Lett.* **333–334**, 211–225 (2012).

88. Lasbleis, M. & Deguen, R. Building a regime diagram for Earth's inner core. *Phys. Earth Planet. Int.* doi:10.1016/j.pepi.2015.02.001 (2015).

89. Korenaga, J. Urey ratio and the structure and evolution of Earth's mantle. *Rev. Geophys.* **46**, 2007RG000241 (2008).

90. Lythgoe, K., Rudge, J., Neufeld, J. & Deuss, A. The feasibility of thermal and compositional convection in Earth's inner core. *Geophys. J. Int.* **385**, 764–782 (2015).

91. Monnereau, M., Calvet, M., Margerin, L. & Souriau, A. Lopsided growth of Earth's inner core. *Science* **328**, 1014–1017 (2010).

92. Aubert, J., Finlay, C. & Fournier, A. Bottom-up control of geomagnetic secular variation by the Earth's inner core. *Nature* **502**, 219–223 (2013).

93. Olson, P. & Deguen, R. Eccentricity of the geomagnetic dipole caused by lopsided inner core growth. *Nat. Geosci.* **5**, 565–569 (2012).

94. Nakagawa, T. & Tackley, P. Implications of high core thermal conductivity on Earth's coupled mantle and core evolution. *Geophys. Res. Lett.* **40**, 1–5 (2013).

95. Labrosse, S., Hernlund, J. & Coltice, N. A crystallizing dense magma ocean at the base of the Earth's mantle. *Nature* **450**, 866–869 (2007).

96. Ziegler, L. B. & Stegman, D. R. Implications of a long-lived basal magma ocean in generating Earth's ancient magnetic field. *Geochem. Geophys. Geosys.* **14**, 4735–4742 (2013).

97. Stevenson, D. How to keep a dynamo running in spite of high thermal conductivity **Abstract AGU Fall Meeting 2012** (2012).

98. Amit, H. Can downwelling at the top of the Earth's core be detected in the geomagnetic secular variation? *Phys. Earth Planet. Int.* **229**, 110–121 (2014).

99. Lesur, V., Whaler, K. & Wardinski, I. Are geomagnetic data consistent with stably stratified flow at the core-mantle boundary? *Geophys. J. Int.* 929–946 (2015).

100. Buffet, B., & Matsui, H. A power spectrum for the geomagnetic dipole moment. *Earth Planet. Sci. Lett.* **411**, 20–26 (2015).

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Author Information Reprints and permissions information is available at www.nature.com/reprints. The authors declare that they have no competing financial interests. Correspondence and requests for materials should be addressed to Chris Davies. (email: c.davies@leeds.ac.uk).

Symbol	100%Fe	82%Fe-8%O-10%Si	79%Fe-13%O-8%Si	81%Fe-17%O-2%Si
$\Delta\rho$ (gm/cc)	0.24 [17]	0.6 [40]	0.8 [33]	1.0 [33]
c_O^S	–	0.0002 [14]	0.0004 [14]	0.0006 [81]
c_{Si}^S	–	0.0554 [14]	0.0430 [14]	0.0096 [81]
c_O^L	–	0.0256 [14]	0.0428 [14]	0.0559 [81]
c_{Si}^L	–	0.0560 [14]	0.0461 [14]	0.0115 [81]
C_p (J/kg/K)	715 [56] — 800 [53]	–	–	–
γ	1.4 [55] — 1.5 [17,56]	–	–	–
$\Delta S(r_\mathrm{i})$ (k_B)	1.05 [17]	–	–	–
$L(r_\mathrm{i})$ (MJ/kg)	0.75	–	–	–
$T_\mathrm{m}(r_\mathrm{i})$ (K)	6350 [17,20]	5900	5580	5320
$\left.\frac{\mathrm{d}T_\mathrm{m}}{\mathrm{d}P}\right\|_{r_\mathrm{i}}$ (K/GPa)	9.01	9.01	9.01	9.01
$\alpha_T(r_\mathrm{i})$ ($\times10^{-5}$/K)	1.0 [54,56]	-	-	-
$T_\mathrm{a}(r_\mathrm{o})$ (K)	4735 17,20	4290	4105	3910
$\left.\frac{\partial T_\mathrm{a}}{\partial P}\right\|_{r_\mathrm{i}}$ (K/GPa)	6.96	6.25	6.01	5.81
$\left.\frac{\partial T_\mathrm{a}}{\partial r}\right\|_{r_\mathrm{o}}$ (K/km)	-1.15	-1.03	-1.00	-0.96
σ ($\times10^6$ S/m)	1.36 [25], 1.4 [23], 1.86 [26,*]	1.12 [25]	1.11 [25]	1.18 [25]
k (W/m/K)	159 [25], 150 [23], 170 [26]	107 [25]	99 [25]	101 [25]
D_O ($\times10^{-8}$ m^2/s)[25]	-	1.31	1.30	-
D_{Si} ($\times10^{-8}$ m^2/s)[25]	-	0.52	0.46	-
ν[25]	6.9	6.8	6.7	-
α_O^D ($\times10^{-12}$ kg/m^3 s)	–	0.72	0.97	1.11
α_{Si}^D ($\times10^{-12}$ kg/m^3 s)	–	1.19	1.10	40.6
		O	Si	
α_c[46,49]	–	1.1	0.87	
$\left(\frac{\partial\mu}{\partial c_X^l}\right)_{P,T}$ (ev/atom)	–	1.02×10^{10}	1.40×10^{10}	

Table 1: **Core material properties for pure iron and three Fe-O-Si mixtures.** Models are named after the molar concentrations of mixtures of Fe, O, and Si corresponding to the given density jump. Quantities in the first section define the core chemistry model. Numbers in the second section determine the core temperature properties in the third section. The core temperature is assumed to follow an adiabat, denoted T_a, and the melting temperature of the core alloy is denoted T_m. CMB values for transport properties calculated along the corresponding adiabats are given in section four. The CMB radius is denoted $r_\mathrm{o} = 3480$ km, the present-day ICB radius is $r_\mathrm{i} = 1221$ km and k_B is Boltzmann's constant. Where a range is given, numbers in red are used in the core models. $*$: This value was derived at a presumed CMB temperature of 3750 K.

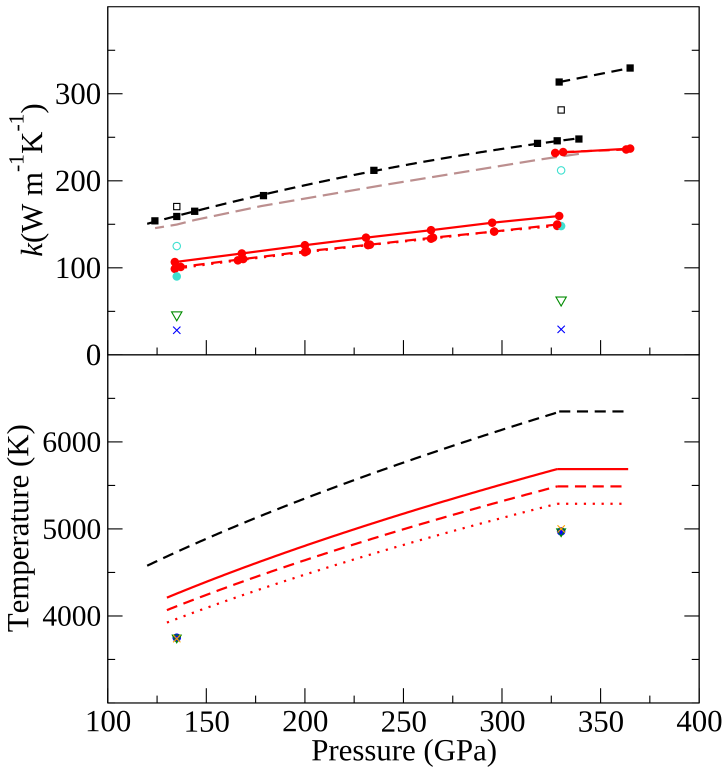

Figure 1: **Comparison of thermal conductivity estimates (top) and adiabatic temperature profiles (bottom) from different studies**. The core chemistry models in Table 1 are shown in black (100%Fe)[24] and red (82%Fe-8%O-10%Si, solid line[25]; 79%Fe-13%O-8%Si, long-dashed line[25]; 81%Fe-17%O-2%Si, short-dashed line[81]). Data from two other recent studies are shown for pure Fe (open black squares[26], brown dashed line[23] using the volume-temperature data of Pozzo et al 2012[24]), a mixture of 76.8%Fe-23.2%O (open aqua circles[26]) and a mixture of 77.5%Fe-22.5%Si (filled aqua circles[26]). Two older estimates of k are shown by the open green triangles[29] and blue crosses[30]. Inner core values were obtained from calculations on solid mixtures[27].

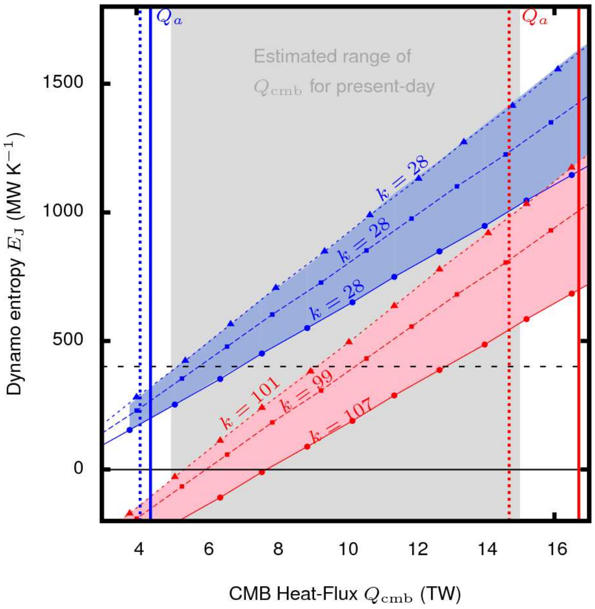

Figure 2: **Present-day core energy budget**. Models shown in red use recent estimates of the thermal conductivity k (red in Figure 1) calculated for ICB density jumps $\Delta\rho = 0.6$ (solid line), 0.8 (long-dashed line) and 1.0 gm/cc (short-dashed line); models in blue all use a low value[30] $k = 28$ W m^{-1} K^{-1}. Other parameters are given in Table 1. Vertical lines indicate ranges for the heat $Q_{\rm a}$ lost down the core adiabat. The black dotted line indicates a plausible estimate[65] for $E_{\rm J}$. Dynamo action requires $E_{\rm J} > 0$. The grey shaded region indicates present-day estimates of CMB heat flow[62,63]. For $Q_{\rm cmb} < Q_{\rm a}$ any convection in the uppermost core is driven compositionally against thermal stratification.

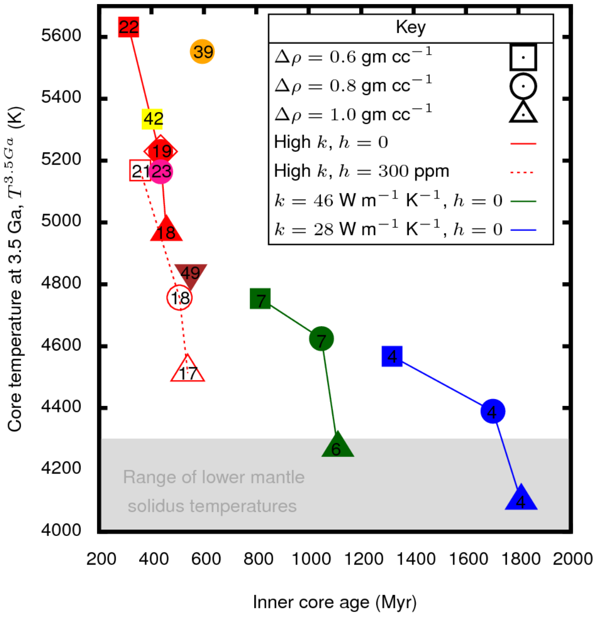

Figure 3: **Core thermal evolution**. Numbers inside each symbol give CMB heat flow (TW) at 3.5 Ga. High k models use the red profiles in Figure 1 that have been calculated for each $\Delta\rho$; models in blue and green use the same k for each $\Delta\rho$. Models joined by lines use $E_J = 0$ prior to inner core formation, after which $Q_{\rm cmb}$ is set constant to ensure the outer core remains just superadiabatic. Results from other recent studies are shown in yellow[68], pink[63], orange[80] and maroon[69]. The inverted triangle denotes that $\Delta\rho$ did not enter into this formulation. Open diamond denotes the reference case in Figure 4.

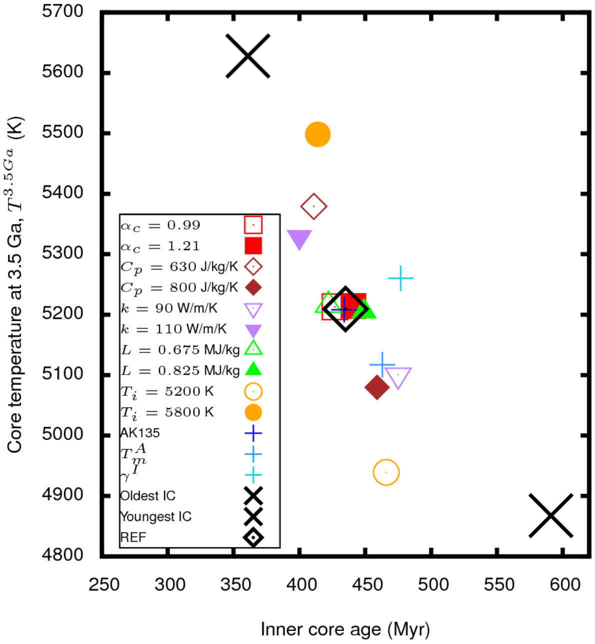

Figure 4: **Dependence of core thermal history predictions on various material properties**. Each model uses identical parameters to the reference model, denoted REF and shown with a red diamond in Figure 3, except the quantity referred to in the legend. A ±10% variation in α_c, C_p and L from the values in Table 1 has been assumed. Values of k refer to the CMB and span the range in Table 1. Values of T_i span the ±300 K error estimates[14,20] described in the text. AK135 is a model of core density[71]. T_m^A is a recent experimentally-determined melting curve[20]. Model γ^I uses depth-dependent γ taken from a recent study[54]. Crosses show the youngest and oldest inner core ages that can be achieved by combining the other variations.