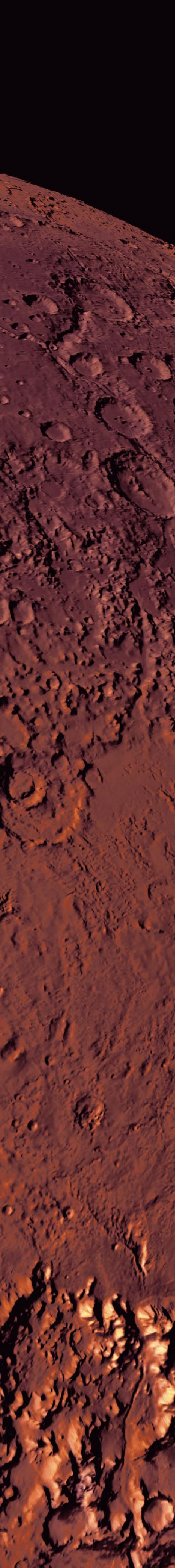

Mars' core and magnetism

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The detection of strongly magnetized ancient crust on Mars is one of the most surprising outcomes of recent Mars exploration, and provides important insight about the history and nature of the martian core. The iron-rich core probably formed during the hot accretion of Mars ~4.5 billion years ago and subsequently cooled at a rate dictated by the overlying mantle. A core dynamo operated much like Earth's current dynamo, but was probably limited in duration to several hundred million years. The early demise of the dynamo could have arisen through a change in the cooling rate of the mantle, or even a switch in convective style that led to mantle heating. Presently, Mars probably has a liquid, conductive outer core and might have a solid inner core like Earth.

Although the existence of the martian core has been accepted for many decades, is interesting for several reasons. First, its size and composition tell us about Mars as a whole — its constituents and provenance. Second, its antiquity tells us about early conditions on Mars; we believe that the core formed early, and this requires that Mars had a hot beginning. Third, this core is the likely source of a magnetic field for some part of Mars' history, probably the earliest part, just as Earth's core is the source of the current geomagnetic field. Fourth, the field may have influenced the early climate through its influence on atmospheric escape. It could also have affected the environment for early life on Mars. Fifth, the heat flow from the core may have fed mantle plumes and influenced volcanic activity, much as hot spots such as Hawaii are thought to be fed by core heat flow on Earth. Sixth, a core, if partly or entirely liquid, influences rotational dynamics, just as (for example) changes in length of day are influenced by Earth's liquid core.

Mars is built from roughly the same ingredients as Earth: silicates and oxides of magnesium and iron, as well as metallic iron (alloyed with various constituents). The mantle and crustal components are discussed in the accompanying article by Zuber (pages 220–227). When we refer to a core for the terrestrial planets Mercury, Venus, Earth and Mars, we mean a central region that is rich in metallic iron. Because this material is about twice as dense as the silicates and oxides making up the crust and mantle, its presence as a core is revealed through its influence on the mean density of the planet and through its effect on the moment of inertia. Old measurements of gravity and more recent geodetic data from Pathfinder[1] reveal that the mean moment of inertia for Mars is $0.365MR^2$, where M is the mass of Mars and R is its mean radius. Together with the martian mean density of 3.93 g cm^{-3}, this suggests a model of Mars that is not too different from a scale model of Earth (that is, similar ingredients, distributed similarly). The martian core is proportionately a little smaller than Earth's core, and proportionately more iron is found in the mantle (in oxides or silicates). As we do not know the composition of the core we cannot be certain about core size, but a core radius of around 1,300–1,500 km (depth to core of 1,900–2,100 km) is indicated. Figure 1 shows a simple interior structure of Mars. Bertka and Fei[2] suggest a composition for Mars that is different from partly devolatilized primitive meteorites.

Far less is known about the martian core than Earth's core because we lack seismological evidence or geodetic data of sufficient precision. In particular, we do not know whether the core is entirely liquid, partly liquid (like Earth) or entirely solid, although there are indirect arguments against an entirely solid core. Fortunately, because Mars is at lower pressures than Earth, and so is more accessible to high-pressure experiment, it is possible to assess the likely phase composition for the core.

I begin with a discussion of the timing of core formation and the new evidence on the nature and origin of martian magnetism. An assessment of the dynamo process follows and is applied to the possible thermal histories of Mars. I conclude with other implications of martian magnetism and core structure and some comments on future exploration.

Core formation

It is widely accepted that terrestrial planetary cores owe their existence to a process of gravitational separation of mostly liquid, immiscible iron from the (partly) solid silicates. The supporting arguments are partly physical[3], but increasingly geochemical. Although we have no samples of either the core of Earth or the core of Mars, we do have rocks that are probably indicative of mantle composition. For Mars, these are the very limited yet highly important SNC meteorites (for shergottites, nakhilites and chassignites). As on Earth, these igneous rocks show a striking depletion of 'iron-loving elements' (called siderophiles) whose extraction testifies to the conditions of core formation[4]. Isotopic data[5,6] also suggest that this core-forming event was early in Mars' history.

These data, together with physical modelling, suggest a scenario similar to the following. Mars accumulated from smaller bodies over a period of perhaps as long as 100 million years (Myr), but possibly much shorter, around 4.5–4.6 billion years (Gyr) ago[7–9]. Isotopic evidence is compatible with a very short accretion time, suggesting that Mars might even have been a runaway, isolated embryo rather than a slowly accumulated body like Earth. In this accretional process, the impacting bodies may have already had iron cores, but the energetics of the impact events would have caused extensive melting and mixing of the immiscible metallic iron and silicate/oxide components, allowing chemical re-equilibration on a small scale (centimetres to metres)[3]. A substantial mass of the impacting bodies may have been in the form of giant impacts (bodies of the order of the mass of Earth's moon),

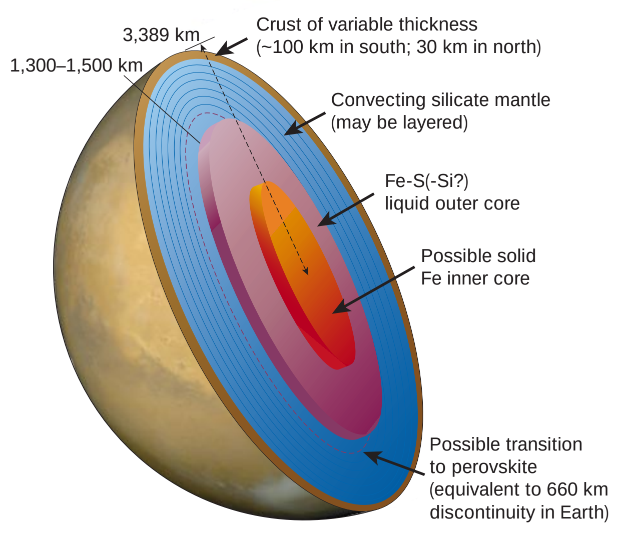

Figure 1 Cutaway view of the martian interior.

but even much smaller bodies can bury a great deal of heat at depth. The energy of gravitational formation of Mars is roughly $0.6GM/R$ per unit mass, where G is the gravitational constant. If all this were converted into heat it would be sufficient to heat the Mars-forming material to several thousand degrees above the melting point, and even with the loss of heat by radiation, a magma ocean is likely. This ocean might be transient (surviving for a brief period after each giant impact), or it might be sustained by a dense steam atmosphere[10], but in either case it will define the conditions of most of the core formation.

Although not as hot or at such high pressure as the likely conditions that formed Earth's core[4], the lower gravity on Mars would still permit a thick magma ocean. Metallic iron can settle as droplets in a convecting magma ocean, to accumulate as large blobs ('diapirs') that then ascend by Stokes flow through the possibly more viscous and high-pressure deep mantle. In this scenario, the core might initially be either a few hundred degrees hotter than the mantle[11] (if the energy of core formation is retained substantially within the iron) or the same temperature as the deep mantle (if efficient thermal equilibration takes place). Much of the martian crust (particularly that preserved in the south) may have formed in this very earliest epoch. Constraints on the timing of crustal formation and thickness are discussed by Zuber (pages 220–227).

Martian magnetism

Mars, unlike Earth, has no global dipole magnetic field. The Mars Global Surveyor spacecraft confirmed this, but also found strong, spatially variable magnetic fields at altitudes of ~200 km down to closest approach of ~110 km (refs 12, 13). Figure 2 shows hemispherical maps of the radial field normalized to a constant 200-km altitude. The fields are measured below much of the martian ionosphere and much of the power in their spatial variability is at length scales comparable to the distance from the surface. 'Inversion' of these data is non-unique, but the source of the field must be confined to the outermost several tens of kilometres of the crust (and possibly confined to an even thinner layer). A deeper layer or source of currents could not provide the observed spatial structure, except with physically implausible assumptions. The inferred crustal magnetizations are up to ~10–30 A m^{-1}, an order of magnitude higher than the strongest magnetizations typically encountered in Earth rocks, and even these values are underestimates if one were to require thinner magnetized layers or incoherent magnetization directions. Given the large amounts of magnetized crust required, it seems very likely that the magnetization is thermal remanence acquired during the last time the rocks cooled through the blocking temperature, for example, following dike injection[14]. Most of these cooling events took place at a time when a large global field was present. ('Most' rather than 'all' is appropriate here, because the surface magnetizations are so large that it is possible for crustal rock to be substantially magnetized through cooling in the presence of other crustal fields, rather than a global field.) Other origins of the magnetization (for example, due to impact, as suggested for the Moon) are conceivable in principle, but seem insufficient given the magnitude of the requirements.

There is great interest in the meaning of the spatial pattern of magnetization, including possible lineations that suggest an analogy to plate tectonically derived lineations of magnetization on Earth's ocean floor, but the current constant-altitude (that is, constant-resolution) maps do not provide strong support for these speculations. Models of the crustal magnetization suggest that the martian field may have undergone reversals.

It seems likely that Mars requires at least one (and preferably several) of the following: high abundance of appropriate magnetic materials (for example, magnetite), a particularly favourable magnetic mineralization (for example, single domains), large volumes of crust that are coherently magnetized, and/or an unusually large field in which the magnetization was acquired.

The fact that Mars did have a global magnetic field for one or more periods in its early history suggests that it once had an active core dynamo, the process responsible for Earth's current field. The strongest magnetizations are observed in the ancient southern highlands of Mars, which predate 4 Gyr. The antiquity of these regions is inferred by the (imprecise) method of crater counts (see review in this issue by Zuber, pages 220–227). However, not all ancient crust on Mars produces large magnetic fields at the spacecraft altitude, and not all younger crust is devoid of magnetization. This prevents firm conclusions being made about the timing of acquisition of magnetization and hence the timing of a postulated martian core dynamo.

Schubert *et al.*[15] have suggested a later (post ~4.0 Gyr) period of magnetization, which would indicate a later period of dynamo activity. But some arguments point towards ancient (4.2 Gyr or earlier) acquisition. First, as noted by Acuna *et al.*[12], the ancient impact structure Hellas (believed to be at least 4 Gyr old) lacks any magnetic signature, and seems to be surrounded by a region with very little coherent magnetization. This is a plausible outcome were the impact to have occurred when Mars possessed no global magnetic field. It is not a plausible outcome if the southern crust were subsequently reheated and then cooled to acquire magnetization during a later epoch in which a global field was active. Second, it is difficult to imagine any physically plausible scenario in which large provinces of the southern crust were extensively heated later in Mars' history without producing some surficial difference in appearance from those regions that were not so treated. This argument is supported by the recognition that huge volumes of crust are required to explain the observed magnetization, rather than some thin layer of possibly remagnetized material. In particular, revived igneous activity generating new crust would certainly disrupt these terrains because of the large amounts required. Third, evidence from the ancient martian meteorite ALH84001 suggests that its magnetization was acquired at 4.0 Gyr or even earlier[16].

A region that lacks large magnetic fields at the spacecraft altitude might still consist of crust that formed in the presence of a global magnetic field. For example, the magnetization may be spatially incoherent, the cooling history may have favoured multidomain magnetite or less favourable mineralization, or the field may have been reversing more rapidly. Moreover, the early rapid pace of planetary evolution means that regions in the south that seem to be of the same age may nonetheless differ in age by ~100 Myr and thus cooled in a different magnetic field, even though their surface

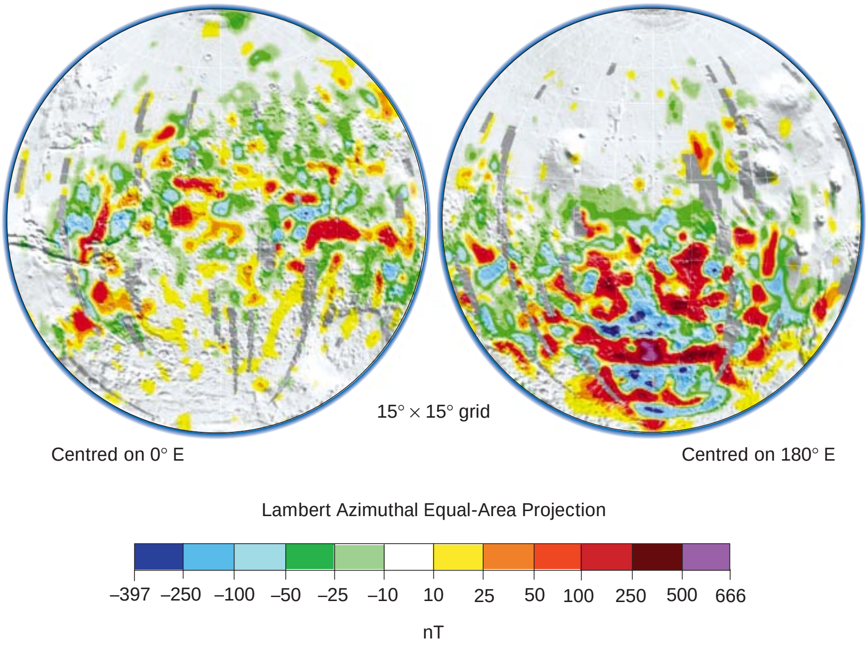

Figure 2 Radial magnetic field at 200-km altitude, based on data collected by the Mars Global Surveyor spacecraft. The southern hemisphere exhibits the largest field anomalies. Notice that the opposite hemisphere has field anomalies that are typically an order of magnitude smaller and exhibit less coherent spatial structure. These maps were prepared by M. Purucker, Goddard Space Flight Center. They are in Lambert Azimuthal Equidistant projection and were published (in a different projection) in ref. 39.

appearance and crater density seem to be identical. Last but not least, ancient crust may underlay younger crust in some northern localities, thereby allowing preservation of a (relatively weak and patchy) magnetization, even when the surface age postdates any global field.

A martian dynamo?

The dynamo mechanism (see Box 1) is much studied but still imperfectly understood[17,18], despite recent advances in numerical simulation[19]. In particular, we do not know the conditions sufficient for the existence of a planetary dynamo. Because we can only speculate about early conditions on Mars, the problem of inferring or predicting the history of a martian dynamo is indeed formidable. Earth's dynamo is also imperfectly understood, although it is thought that it arises from convection driven largely by inner-core growth[20–22].

If a dynamo exists, then it is likely that the expected field magnitude B inside the region of field generation is given by the Elsasser number of order unity. This implies $B \sim (2\rho\Omega/\sigma)^{1/2}$, where ρ is the fluid density, Ω is the planetary rotation rate and σ is the electrical conductivity. For the martian core, this yields $\sim 10^{-3}$ tesla, but because this is the same prediction as for present Earth, the predicted palaeofield at the surface of Mars is indistinguishable (at the level of this crude argument) from the present field at Earth's surface. Dynamo theory admits weaker fields as possible solutions, but it does not admit fields substantially larger than $B \sim (2\rho\Omega/\sigma)^{1/2}$. Despite suggestions to the contrary[22], there is no theoretical basis at present for the idea that the field scales in some direct way with the energy source, so that it might undergo slow decline over geological time or large changes arising from inner-core nucleation.

One speculative explanation for the origin of magnetization on Mars is that the field was generated in a magma ocean. Plausible numbers are a characteristic fluid velocity v of $\sim 10^{-1}$ m s^{-1} (because of very high heat flows at that time), a characteristic length scale L of $\sim 10^6$ m, and a magnetic diffusivity λ of 10^4 m^2 s^{-1} (possibly appropriate to high-temperature and high-pressure silicate melts[23]), which together give a magnetic Reynolds number R_m of ~10. This is marginal at best, but would be attractive because large fields are predicted (~0.01–0.1 tesla at the martian surface). The extremely high observed magnetizations might then be explained, although a core dynamo is more plausible.

Thermal or compositional convection?

If one accepts that core convection is needed, then a probable necessary condition for a dynamo is the presence of convection. In terrestrial planets (including Earth), the criterion for core convection is difficult to satisfy. The reason for this is that the natural scale for core heat flows is such that this heat can probably be carried by conduction at a temperature gradient that is stably stratified (that is, it inhibits convection). To obtain core convection, one must appeal to unusually large heat flows or the development of an inner core. In either case, the core must be cooling. To appreciate this argument, consider first the simple case of no inner core. Convection will occur provided the heat flux within the core exceeds that which is carried by conduction along an adiabat:

$$F_{\mathrm{total}} > F_{\mathrm{cond,ad}} \equiv k\alpha Tg/C_p \Leftrightarrow \text{thermal convection}$$

where k is the thermal conductivity, α is the coefficient of thermal expansion, T is the temperature, g is the gravitational acceleration and C_p is the specific heat at constant pressure. These parameters are all slowly varying within a core (if T is close to being adiabatically distributed), except for g, which is approximately linear in radius r, the distance from the planet centre. If the core is simply cooling and releasing the stored sensible heat (provided by gravity during planetary accretion), then the total heat flux is also linear in r: $F_{\mathrm{total}}(r) = -\rho C_p r(\mathrm{d}T_c/\mathrm{d}t)/3$, where T_c is the mean core temperature and t is time. It is unlikely that the core contains significant radioactive heat sources (even less likely than Earth, where one can always appeal to unknown, very high pressure effects). Consequently, if thermal convection ceases to operate in the outer part of the core, then it will also cease to operate at about the same time elsewhere in the core.

If the core is cooling and the central temperature drops below the liquidus for the core alloy, then an inner core will nucleate. In Earth, we know from seismic evidence that the core is ~10% less dense than pure iron, and many suggestions have been offered for the identity of the light elements that are mixed with the iron[24,25]. At the lower pressures relevant to Mars, the dominant light element may be sulphur[2]. For sulphur abundances that are less than cosmic relative to iron, as cosmochemical arguments suggest, the inner core will be nearly pure iron (with some nickel) and the sulphur will be entirely in the outer core. The introduction of this light element into the fluid of the lowermost core will tend to promote convection and cause mixing throughout all or most of the outer core, provided the cooling is sufficiently fast. Latent heat release at the inner-core/outer-core boundary will also contribute to the likelihood of convection. However, inner-core growth permits outer-core convection even when the heat flow through the core–mantle boundary is less (perhaps much less) than the heat carried by conduction along an adiabat. In this regime (possibly that occupied by Earth), the temperature gradient is very slightly less steep than adiabatic and the compositional convection carries heat downwards. The total heat flux is still outwards, of course, as the heat carried by conduction is large. This state is possible because the buoyancy release associated with the compositional change exceeds the work done against the unfavourable thermal stratification. Unlike thermal convection, compositional convection may not cease everywhere throughout the core at a single epoch. This argument is modified in detail but not in general outline, should the core include a light element that does not exhibit eutectic behaviour (for example, silicon), as well as the (universally expected) complement of sulphur.

Required cooling rates

For plausible choices of parameters, the cooling rate of the core must exceed about 80 K Gyr^{-1} to obtain thermal convection. However, this estimate is uncertain by perhaps as much as a factor of two. The required cooling rate in the presence of a growing inner core is much smaller, by a factor of several[26,27], but has not been studied in detail for Mars. As a consequence, models with an inner core will tend to sustain a dynamo for a long time, perhaps even to the present day, unless there is something unusual about the thermal history (as suggested below). The overlying mantle determines the cooling rate. Indeed, it is the mantle that determines whether a terrestrial planet has core convection and whether it can have a dynamo.

It is also possible that the sufficient condition for a dynamo is not far removed from the necessary condition for the presence of any convection[26]. Simple scaling laws for convection (compatible with the philosophy of Kolmogorov turbulence and known to astrophysicists as mixing length theory) suggest that $v \sim 0.1(F_{\text{conv}}/\rho)^{1/3}$, where ρ is the fluid density and F_{conv} is the convective heat flux (or its compositional equivalent when the convection is driven by compositional density differences). I define

$$\varepsilon = F_{\text{conv}}/F_{\text{cond,ad}} = (F_{\text{total}} - F_{\text{cond,ad}})/F_{\text{cond,ad}}$$

Substitution above shows immediately that for plausible parameters in the martian core ($L \sim 10^6$ m, $\lambda \sim 1\ \text{m}^2\,\text{s}^{-1}$, $\rho \sim 10^4\ \text{kg}\,\text{m}^{-3}$), R_{m} may be large even if $\varepsilon \ll 1$. That is, the heat flow has to only slightly exceed that for any convection in order to reach that for convection of sufficient vigour to sustain a dynamo. This claim must be tested by further numerical work. It is conceivable, but difficult energetically, for a dynamo to function for $\varepsilon < 0$ (for example, because of baroclinic instabilities and thermal winds arising from horizontal temperature gradients that are caused by lateral differences in heat flow through the core–mantle boundary). Even in this case, one would expect that a dynamo requires $|\varepsilon| \ll 1$, as the vertical motions would otherwise be strongly suppressed and this inhibits dynamo activity. Alternatives to convective driving (for example, precession[28]) still require the core to be close to adiabatic and thus do not escape the constraints discussed above.

In conclusion, if the mantle cools fast enough (or is cool enough to allow inner-core nucleation) then a dynamo occurs, but if the mantle is too hot or fails to cool then there is no dynamo.

Possible histories of the martian core

Three possible scenarios for the history of the martian dynamo are presented in Figure 3. The first is the simplest: the planet starts out very hot and cools quickly at first. The core remains completely liquid throughout. As the cooling rate declines, a point is reached at which the heat flow out of the core can be accommodated by conduction alone. At that epoch, the dynamo turns off (in a very short time geologically, perhaps as little as a few thousand years) and no further field generation is possible, provided an inner core never develops. This model requires that the core of Mars is sulphur-rich, perhaps 10% or more by mass. It also requires tuning of the parameters so that the dynamo turns off as early in Mars' history as some arguments suggest. Most published models are of this kind[22,29].

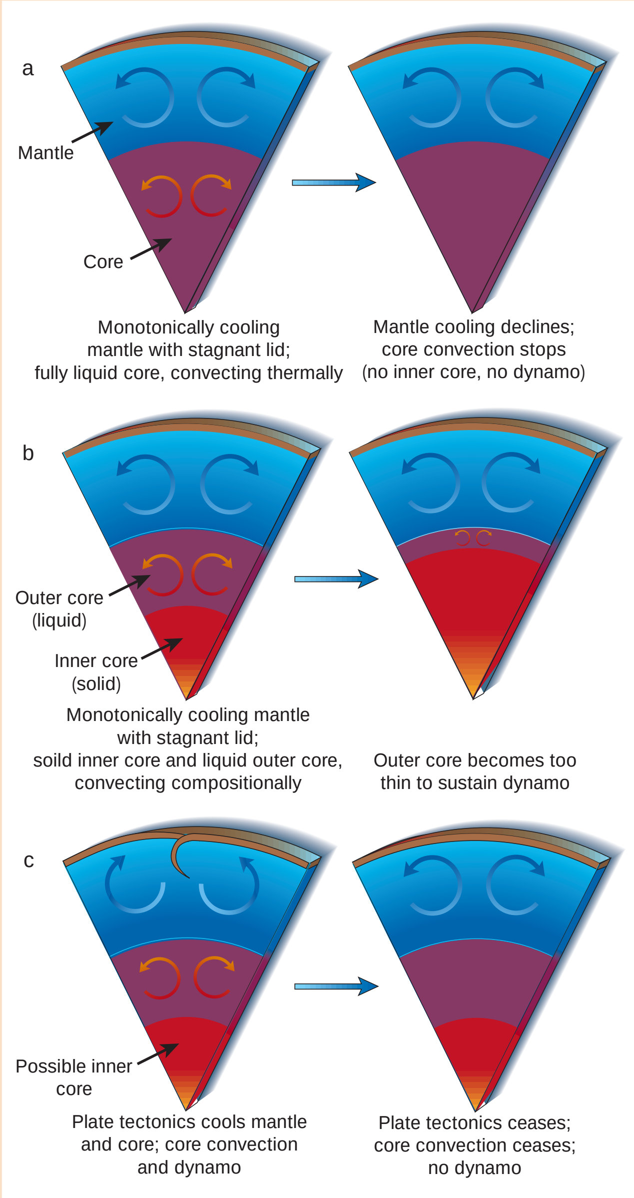

Figure 3 Possible scenarios of martian core evolution. **a**, High-sulphur model; **b**, low-sulphur model; **c**, plate-tectonic model (works for all sulphur contents).

The second scenario is almost the antithesis of the first. It has not been modelled in detail, although it is implicit in the early work of Young and Schubert[30], who considered the possibility of complete core freezing. In this model, the sulphur content for the core is sufficiently low that an inner core develops early and grows rapidly. The liquid outer core becomes progressively more sulphur-rich and evolves towards the eutectic composition. The experimental data[31] yield a eutectic of around 1,400 K at the top of the core. For realistic models of Mars' mantle convection[32], the expected present-day core–mantle boundary temperature is at least 1,850 K, so there is no

Box 1
What is a dynamo?

The essence of a dynamo lies in electromagnetic induction — the creation of currents and associated field through the motion of conducting fluid across magnetic field lines. Numerical and analytical work indicate that a dynamo will exist if the fluid motions have certain desired features and the magnetic Reynolds number R_m exceeds about 10. Here,

$$R_m \equiv vL/\lambda$$
$$\lambda \equiv 1/\mu_0\sigma$$

where v is a characteristic fluid velocity, L is a characteristic length scale of the motions or field (for example, the core radius), λ is called the magnetic diffusivity, μ_0 is the permeability of free space and σ is the electrical conductivity (SI units). It seems likely that fluid motions of the desired character arise naturally in a convecting fluid (irrespective of the source of fluid buoyancy), provided the Coriolis force has a large effect on the flow, that is, $v/\Omega L < 1$, where Ω is the planetary rotation rate. This is easily satisfied for any plausible fluid motion of interest. Realistic dynamos often require a somewhat larger value than ~10 for R_m, especially if driven by secular cooling without an inner core[40], and attention must also be paid to the different kinds of motions of relevance (for example, vertical motions and differential rotation may have different amplitudes and scales of variation).

prospect that the outer core will completely freeze. However, it is conceivable that the outer core will become sufficiently thin that dynamo activity can no longer be sustained. This would seem implausible based on simple scaling arguments, but it might be the state that Mercury currently occupies[33]. It probably requires a lower sulphur content of the martian core than most would consider plausible, perhaps no more than a few per cent (even less than typical estimates for Earth). Further dynamo simulations are needed to test this hypothesis.

The third scenario invokes a change in mantle convection to trigger the death of the martian dynamo[32]. It is assumed that early Mars had mobile lid convection in which the lithosphere could be recycled. On Earth, this is accomplished by plate tectonics, and this could also be the case on Mars[34]. (It is, however, the recycling of the lithosphere that matters, not the form of the recycling; so there is no need to assume that Mars did exactly what Earth does.) At some time, perhaps after only a few hundred million years, this process ceased and Mars evolved slowly into the stagnant lid regime that it (and all terrestrial bodies except Earth) currently occupies. If this regime follows one of lithospheric recycling, then the mantle must heat up, because the elimination of heat is less efficient. In other words, the coldest time for the martian mantle was early in Mars' history, despite the inexorable monotonic decline of radioactive heat sources in the mantle and the crust. This scenario has the advantage that it may work for all possible sulphur contents in the core, as the presence of an inner core will not drive a dynamo if the mantle minimum temperature was reached early in Mars' history. An inner core drives a dynamo only while it is growing, and it can grow only if the core is cooling. One problem with this scenario is that it invokes an *ad hoc* timing for the cessation of 'plate tectonics'; it also implies the ability for Mars to be volcanically active throughout geological time.

In all these scenarios, the beginning of dynamo activity may be delayed after Mars' accretion until a thermal boundary layer builds up in the lowermost mantle, depending on the uncertainty in the initial temperature difference between the core and mantle. But this is unlikely to produce a delay of more than ~100 Myr.

There may be other scenarios not yet considered. Unfortunately, none of these scenarios can be tested with great confidence because the parameters that define their chronologies are not known with sufficient accuracy. However, the presence or absence or size of the inner core is clearly a crucial variable and may eventually be determined by a combination of geodesy and seismology. Numerical dynamo modelling will also be important in the coming years.

Consequences of the martian core and dynamo

Core cooling dictates the presence of a thermal boundary layer at the base of the overlying mantle. Plumes can detach from this layer and may be a cause of hot-spot volcanism. Harder and Christensen[35] have proposed that Mars may be in a regime where a single plume dominates because of the effect of a major endothermic phase transition near the base of the martian mantle (the same phase transition that defines the upper-mantle/lower-mantle boundary on Earth). This plume might be stable for a long period of time, perhaps billions of years, and may be responsible for the Tharsis volcanic province. This hypothesis provides the exciting prospect of linking core thermal history with martian volcanic history. However, it leaves unanswered several questions. If the core heat flow is so low (as required by the absence of a dynamo throughout much of Mars' history), then is it reasonable to suppose that it is responsible for the dominant volcanic activity on Mars? Why would a deep-seated plume happen to produce volcanism at a location just northward of the principal geological feature (the crustal dichotomy)? Why is the plume so stable? Perhaps the answer to Tharsis lies nearer the surface of Mars rather than in the core history.

The history of the atmosphere[36] may also be influenced by the magnetic field history through the effect of the field on atmospheric sputtering. The history of martian magnetism might even be linked to the history of life on Mars. Perhaps the strongest argument for a biological effect in ALH84001 lies in the single-domain magnetite grains[37], whose presence in biological organisms is useful only while Mars has a field. This might also push martian magnetism back to the earliest epoch.

The future

Although martian core studies can benefit from work in all areas of planetary science (including geochemistry), the greatest contribution is likely to arise from seismological and geodetic efforts. In particular, the Mars Netlander mission[38] and subsequent follow-ups are likely to have the greatest role. It may also be essential to better characterize the surface magnetization, something that no currently funded mission can do. We can also look forward to exciting developments in our understanding of dynamos. Mars' core is at least as interesting as Earth's core for our general understanding of planet evolution. □

1. Folkner, W. N., Yoder, C. F., Yuan, D. N., Standish, E. M. & Preston, R. A. Interior structure and seasonal mass redistribution of Mars from radio tracking of Mars Pathfinder. *Science* **278,** 1749–1752 (1997).
2. Bertka, C. M. & Fei, Y. W. Implications of Mars Pathfinder data for the accretion history of the terrestrial planets. *Science* **281,** 1838–1840 (1998).
3. Stevenson, D. J. in *Origin of the Earth* (eds Newsom, H. E. & Jones, J. E.) 231–250 (Oxford Univ. Press, New York, 1990).
4. Righter, K., Hervig, R. L. & Kring, D. A. Accretion and core formation on Mars: molybdenum contents of melt inclusion glasses in three SNC meteorites. *Geochim. Cosmochim. Acta* **62,** 2167–2177 (1998).
5. Chen, J. H. & Wasserburg, G. J. Formation ages and evolution of Shergotty and its parent planet from U-Th-Pb systematics. *Geochim. Cosmochim. Acta* **50,** 955–968 (1986).
6. Lee, D. C. & Halliday, A. N. Core formation on Mars and differentiated asteroids. *Nature* **388,** 854–857 (1997).
7. Wetherill, G. W. Provenance of the terrestrial planets. *Geochim. Cosmochim. Acta* **58,** 4513–4520 (1994).
8. Chambers, J. E. & Wetherill, G. W. Making the terrestrial planets: N-body integrations of planetary embryos in three dimensions. *Icarus* **136,** 304–327 (1998).
9. Agnor, C. B., Canup, R. M. & Levison, H. On the character and consequences of large impacts in the late stage of terrestrial planet formation. *Icarus* **142,** 219–237 (1999).
10. Matsui, T. & Abe, Y. Formation of a magma ocean on the terrestrial planets due to the blanketing effect of an impact-induced atmosphere. *Earth Moon Planets* **34,** 223–230 (1986).
11. Flasar, F. M. & Birch, F. Energetics of core formation: a correction. *J. Geophys. Res.* **78,** 6101–6103 (1973).
12. Acuna, M. H. *et al.* Global distribution of crustal magnetization discovered by the Mars Global Surveyor MAG/ER experiment. *Science* **284,** 790–793 (1999).
13. Connerney, J. E. P. *et al.* Magnetic lineations in the ancient crust of Mars. *Science* **284,** 794–798 (1999).
14. Nimmo, F. Dike intrusion as a possible cause of linear Martian magnetic anomalies. *Geology* **28,** 391–394 (2000).

15. Schubert, G., Russell, C. T. & Moore, W. B. Timing of the martian dynamo. *Nature* **408,** 666–667 (2000).
16. Weiss, B. *et al.* Records of an ancient martian magnetic field in ALH84001. *Nature* (submitted).
17. Merrill, R. T., McElhinney, M. W. & McFadden, P. L. *The Magnetic field of the Earth* (Academic, New York, 1998).
18. Busse, F. H. Homogeneous dynamos in planetary cores and in the laboratory. *Annu. Rev. Fluid Mech.* **32,** 383–408 (2000).
19. Roberts, P. H. & Glatzmaier, G. A. Geodynamo theory and simulations. *Rev. Mod. Phys.* **72,** 1081–1123 (2000).
20. Loper, D. E. Some thermal consequences of a gravitationally powered dynamo. *J. Geophys. Res.* **83,** 5961–5970 (1978).
21. Gubbins, D., Masters, T. G. & Jacobs, J. A. Thermal evolution of the Earth's core. *Geophys. J. R. Astron. Soc.* **59,** 57–99 (1979).
22. Stevenson, D. J., Spohn, T. & Schubert, G. Magnetism and thermal evolution of the terrestrial planets. *Icarus* **54,** 466–489 (1983).
23. Tyburczy, J. A. & Fisler, D. K. in *Mineral Physics and Crystallography. A Handbook of Physical Constants* (ed. Ahrens, T. J.) 185–208 (Am. Geophys. Union, 1995).
24. Li, J. & Agee, C. B. Element partitioning constraints on the light element composition of the Earth's core. *Geophys. Res. Lett.* **28,** 81–84 (2001).
25. Gessmann, C. K. , Wood, B. J. , Rubie, D. C. & Kilburn, M. R. Solubility of silicon in liquid metal at high pressure: implications for the composition of the Earth's core. *Earth Planet. Sci. Lett.* **184,** 367–376 (2001).
26. Stevenson, D. J. Planetary magnetic fields. *Rep. Prog. Phys.* **46,** 555–620 (1983).
27. Lister, J. R. & Buffett, B. A. The strength and efficiency of thermal and compositional convection in the geodynamo. *Phys. Earth Planet. Int.* **91,** 17–30 (1995).
28. Stevenson, D. J. Planetary magnetism. *Icarus* **22,** 403–415 (1974).
29. Schubert, G. *et al.* in *Mars* (eds Kieffer, H. H., Jakosky, B. M., Snyder C. W. & Matthews, M. S.) 147–183 (Univ. Arizona Press, Tucson, 1992).
30. Young, R. E. & Schubert, G. Temperatures inside Mars: is the core liquid or solid? *Geophys. Res. Lett.* **1,** 157–159 (1974).
31. Fei, Y. W., Bertka, C. W. & Finger, L. W. High-pressure iron-sulfur compound, Fe_3S_2, and melting relations in the Fe-FeS system. *Science* **275,** 1621–1623 (1997).
32. Nimmo F. & Stevenson, D. J. Influence of early plate tectonics on the thermal evolution and magnetic field of Mars. *J. Geophys. Res.* **105,** 11969–11979 (2000).
33. Schubert, G., Ross, M. N., Stevenson, D. J. & Spohn, T. in *Mercury* (eds Chapman, C. *et al.*) 429–460 (Univ. Arizona Press, Tucson, 1988).
34. Sleep, N. H. Martian plate tectonics. *J. Geophys. Res.* **99,** 5639–5655 (1994).
35. Harder, H. & Christensen, U. R. A one-plume model of martian mantle convection. *Nature* **380,** 507–509 (1996).
36. Brain, D. A. & Jakosky, B. M. Atmospheric loss since the onset of the Martian geologic record: combined role of impact erosion and sputtering. *J. Geophys. Res.* **103,** 22689–22694 (1998).
37. Thomas-Keprta, K. L. *et al.* Truncated hexa-octahedral magnetite crystals in ALH84001: presumptive biosignatures. *Proc. Natl Acad. Sci. USA* **98,** 2164–2169 (2001).
38. Special Issue. *Planet. Space Sci.* **48,** 1143–1420 (2000).
39. Purucker, M. *et al.* An altitude-normalized magnetic map of Mars and its interpretation. *Geophys. Res. Lett.* **27,** 2449–2452 (2000).
40. Kutzner, C. & Christensen, U. Effects of driving mechanisms in geodynamo models. *Geophys. Res. Lett.* **27,** 29–32 (2000).