8.09 Magnetic Polarity Reversals in the Core
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8.09.1 Observations

The observational evidence for evaluating and improving geodynamo simulations comes from direct observations of the geomagnetic field over the past few centuries and from the paleomagnetism of magnetic materials, such as rocks, fired artifacts, adobe, and even pigments from paintings, over the rest of geologic time (see Chapters 5.01, 5.02, 5.03, 5.04, 5.05, 5.06, 5.07, see also: Chapters 5.08, 5.09, 5.10, 5.11, 5.12, 5.13, and 5.14). Thus, for information about most geomagnetic behaviors we rely upon the results of paleomagnetic studies. These show us that the Earth’s magnetic field has been dominantly dipolar in the past, as it is today. Moreover, to a reasonable approximation it averages to axial dipolar over a few tens of thousands of years (e.g., Merrill, 1996). This is demonstrated for the past 150 million years by the agreement between plate reconstructions based on marine magnetic anomalies and paleomagnetic apparent polar wander paths (Besse and Courtillot, 2002). The preponderance of paleomagnetic evidence suggests that dipolar dominance was typical earlier as well (McElhinny, 2004), perhaps even during much of the Pre-Cambrian, but Pre-Cambrian evidence is sparse and equivocal (Dunlop and Yu, 2004). To second order, however, a small but significant departure from dipolar of the time-averaged field has long been noted (Wilson, 1971). Averaged over the past 5 My, the departure can be represented as an axial quadrupole contribution that is 3–5% of, and the same sign as, the axial dipole (Johnson and Constable, 1997; McElhinny, 2004). Comparison of paleomagnetic and plate-tectonic continental reconstructions suggests that a 3% axial quadrupole component may have been typical over the past 200 My (Besse and Courtillot, 2002).

The other salient feature of the field is that it reverses polarity. Geologically speaking, reversals occur very quickly. Even for the most recent reversal (Matuyama–Brunhes), which occurred 0.78 Ma, the dating error of the most favorable basalt flows is comparable to the transition duration (Singer et al., 2005). For this reason transition duration must be obtained from sedimentary records, using the stratigraphic thicknesses of the transition zones and estimates of the deposition rate. From 30 selected records of the last four reversals, Clement (2004) found that the average time for the field to change direction from one polarity to the other was 7000 years (Figure 1). These reversals, so defined, tended to occur more quickly at low latitudes than at mid-to-high latitudes, with individual durations ranging from 2000 to 12 000 years. This regionally varying behavior implies that the nondipole field played a large role.

In detail, reversal records display a wide variety of field behavior (Coe and Glen, 2004). Almost all exhibit large intensity drops during transition, but their directional behavior varies greatly. The most detailed, high-deposition-rate lava-flow records and sedimentary records show that at least some reversals are complex, with episodes of oscillatory and rapid field change (Figures 2 and 3). Often the field reverses briefly, but relapses to intermediate directions one or more times before finally attaining stable opposite polarity. This behavior complicates estimation of duration. For instance, if one includes an early unsuccessful swing to normal polarity in the most recent reversal transition, its duration is about 18 000 years (Figure 4), three times longer than if one considers that swing to be an unrelated precursor (Singer et al., 2005). Inclusion of the precursory swing in direction seems reasonable, in light of the complex transition paths of Figures 2 and 3. Even excluding the precursor, high-resolution sediment records show that this transition is complex, with five or six large swings of the field direction (Figure 5) (Channell and Lehman, 1997). Some dynamo simulations have also produced comparably complex and long reversal transitions – for example, the second reversal during...
the tomographic simulation of Glatzmaier et al. (1999) (see case ‘h’ of Figure 14 and Coe and Glen (2004, figure 7 and plate 1)).

Nonetheless, various authors have discerned some statistical regularity in the directional behavior of reversals over the past 2–20 My: the tendency of transitional virtual geomagnetic poles (VGPs) to cluster in preferred longitudinal bands or patches on Australasia and the Americas and to avoid the Central Pacific Basin (Laj et al., 1991; Clement, 1991; Hoffman, 1992; Hoffman and Singer, 2004; Love, 2000), though the statistical significance of this tendency has been disputed (Valet et al., 1992; McFadden et al., 1993; Prévot and Camps, 1993;
The Gauss–Matuyama (2.58 Ma) reversal record of VGPs recorded in sediments deposited in Searles Lake, California (Glen et al., 1999b). Note the highly complex VGP path, with initial and final excursions in orange, multiple rapid oscillations in black, and main reversing phase including two large swings from high to equatorial latitudes in red.

Figure 4  $^{40}$Ar/$^{39}$Ar ages of 23 transitionally magnetized lava flows from four widely spaced localities that record transitional field directions attributed to the most recent, Matuyama–Brunhes (M–B) reversal. Gray bands show the weighted mean age and 2σ uncertainty for each lava sequence. The ages of the lavas from Maui and the uppermost flow from La Palma correspond to the accepted age of the M–B reversal from sedimentary cores (Tauxe et al., 1996), whereas the others correspond to what has been termed the M–B precursor (P) (Hartl and Tauxe, 1996). The probability density curve indicates that these lavas together span a minimum of 18 ky, a period about three times longer than the conventionally cited M–B duration. Adapted from Singer BS, Hoffman KA, Coe RS, et al. (2005) Structural and temporal requirements for geomagnetic field reversal deduced from lava flows. Nature 434: 633–636.
The long timescale suggests mantle influence, and the preferred areas do correlate with large-scale seismic tomography, overlying regions of higher than average P- and S-wave velocity (Laj et al., 1991; Hoffman and Singer, 2004). Assuming that higher seismic velocity signifies regions with lower than average lower-mantle temperature, convective downwelling in the core could be localized there and concentrate poloidal flux lines. Transitional VGP preference would then be expected (Gubbins and Coe, 1993) and has in fact been produced by dynamo simulations employing appropriate heat-flux boundary conditions at the core–mantle boundary (CMB) (Coe et al., 2000; Olson and Christensen, 2002; Kutzner and Christensen, 2004).

Earth’s reversals occur aperiodically, in fact almost randomly, but the mean duration appears to change progressively over long time intervals (Figure 6). From 0 to 165 Ma the field reversed on average about 2 times per million years, but over 10 My intervals the average reversal rate varied from highs of at least 5 per million year to a low of 0. During the interval from 124 to 83 Ma, the
so-called Cretaceous Normal Superchron (CNS), no true polarity reversals have been unequivocally demonstrated. From analysis of the latitudinal variation of paleomagnetic secular variation recorded by lava flows over the past 150 My, McFadden et al. (1991) proposed that average reversal rate is related to symmetry of the nonaxial dipole field: a higher degree of symmetry of the field about the equator, as expressed by its spherical harmonic gauss coefficients, correlates with more frequent reversals (Figure 6). For example, the equatorial dipolar and axial quadrupolar harmonics of the field are symmetric and the axial octupolar harmonic is antisymmetric (Merrill, 1996). Recent analysis of the Glatzmaier et al.’s (1999) dynamo simulations supports this idea: Figure 7 shows that the most stable, nonreversing simulation (case ‘e’ of Figure 14) has far more energy associated with antisymmetric than with symmetric gauss coefficients (Coe and Glatzmaier, 2006). Furthermore, a

Figure 6  Smoothed reversal rate (thin wavy line) corresponding to normal and reversed chrons (black and white bars) of the geomagnetic polarity timescale. The rate goes to zero in the Cretaceous Normal Superchron, from 83 to 124 Ma. The squares denote $b/a$, a relative measure of the average ratio of the antisymmetric to symmetric parts of the geomagnetic field (excluding the axial dipole) as a function of time in the past, inferred from paleomagnetic secular variation using the results of McFadden et al. (1991). These demonstrate a strong inverse correlation between $b/a$ and reversal rate. We note that recent studies (see Ogg (2004)) suggest even higher reversal rates from 155 to 165 Ma than shown here (~10 per My). Adapted from Coe RS and Glatzmaier GA (2006) Symmetry and stability of the geomagnetic field. Geophysical Research Letters 33: L21311 (doi:10.1029/2006GL027903).

Figure 7  Time-averaged spatial energy density $W_{n,m}$ at the CMB associated with each harmonic degree and order up to $n$, $m = 5,5$ for five of the simulations of Glatzmaier et al. (1999) that are shown in Figure 14. The terms associated with gauss coefficients $g_{n,m}$ and $h_{n,m}$ are antisymmetric about the equator when $(n + m)$ is odd and are symmetric about the equator when $(n + m)$ is even. Note that the most stable case, E, heavily favors antisymmetric energy terms (with $n + m$ odd), as does generally the second most stable case D. Adapted from Coe RS and Glatzmaier GA (2006) Symmetry and stability of the geomagnetic field. Geophysical Research Letters 33: L21311 (doi:10.1029/2006GL027903).
simulation with a solid inner core only one-quarter of its size today (Roberts and Glatzmaier, 2001) produced an even more antisymmetric nonaxial dipole, suggesting that reversals may have been much less common in the distant geologic past when Earth’s inner core was smaller (see Chapter 8.02). Limited paleomagnetic evidence available from rocks older than 1 Ga appears to support this suggestion (Coe and Glatzmaier, 2006).

The CNS was clearly a time of exceptional field stability, though a few instances of short-lived reversed directions have been reported within it (Tarduno, 1990; Gilder et al., 2003; Ogg et al., 2004). These might represent aborted reversal attempts. Much more recently, during the past 0.78 My of today’s normal polarity epoch when global simultaneity is easier to establish, there have been eight or more brief excursions of the field to low intensity and reversed or nearly reversed directions that have been detected around the globe (Figure 8) (Champion et al., 1988; Guyodo and Valet, 1999; Lund et al., 2006). They are generally distinguished from complete polarity reversals by their short duration, a few thousand to at most a few tens of thousands of years. In the paleomagnetic and seafloor magnetic anomaly records their classification is somewhat arbitrary (Acton et al., 2006), but in dynamo simulations aborted reversals are those for which the field deep in the core maintains its original polarity throughout the time of intermediate and reversed directions at the surface of the Earth (Figure 9) (Glatzmaier et al., 1999).

8.09.2 Models

Although the magnetic dipole reversal mechanism is still poorly understood, mathematically it is easy to see why there can be two oppositely directed magnetic basins of attraction. Given a solution to the set of equations that govern the thermodynamics and magnetohydrodynamics (MHD) for a convective dynamo (i.e., fluid flow, magnetic field, and thermodynamic variables), completely reversing the magnetic field everywhere would also satisfy the same set of equations; that is, reversing the sign of the magnetic field vector everywhere will not change the Lorentz force or Joule heating because the field is quadratic in these terms. It will also not affect the magnetic induction or conservation equations because the field, although linear in these two equations, appears in every term.

Reversals of the dipolar part of the magnetic field were originally studied using two-dimensional (2-D, axisymmetric) models of the mean magnetic fields. These models are kinematic, that is, the fluid flow is prescribed or parametrized instead of being part of the solution. Only the longitudinally averaged part of

![Figure 8](image-url) Excursions of the geomagnetic field since the last reversal (Brunhes–Matuyama), as indicated by deep minima in Earth’s dipole moment (VADM). Results are from a global stack of records of relative paleointensity data derived from marine sedimentary cores, normalized to fit volcanic absolute paleointensity results. Adapted from Guyodo Y and Valet JP (1999) Global changes in intensity of the Earth’s magnetic field during the past 800 kyr. Nature 399: 249–252.
the magnetic field is calculated (see Chapter 8.03). This method continues to be employed in the solar community to study the periodic reversals associated with the sunspot cycle (e.g., Bonanno et al., 2006) and in the geophysics community to study the aperiodic paleomagnetic reversals (e.g., Giesecke et al., 2005). Such models prescribe the two main ingredients for a self-sustaining dynamo: the ‘alpha effect’, which twists toroidal (longitudinally directed) magnetic field into poloidal (radially and latitudinally directed) field, and the ‘omega effect’, which shears poloidal field into toroidal field. The alpha effect can also act on poloidal field to generate (or destroy) toroidal field. The alpha prescription in these models parametrizes the effects of helical fluid flow and the omega prescription represents the effects of differential rotation (i.e., the variation of angular velocity with radius and latitude). Replacing these complicated, nonlinear, 3-D, time and spatially dependent processes with 2-D and usually time-independent functions greatly simplifies the problem. These studies formed a basis for understanding convective dynamos. The simplifications and assumptions built into these models greatly reduce the complexity and computational expense and therefore can provide good statistics because of the long times that can be simulated. Unfortunately with this approach one cannot reconstruct what the 3-D dynamics must be that maintains the prescribed alpha and omega effects. Discovering and understanding the details of the dynamo mechanisms requires a self-consistent solution of the full set of nonlinear 3-D equations that represent conservation of mass, momentum, energy, magnetic flux, and magnetic induction (see Chapters 8.05 and 8.08).

Early, self-consistent MHD models of the solar dynamo were developed by Gilman (1983) and Glatzmaier (1985). Instead of prescribing the alpha and omega effects parametrically, they solved a full set of MHD equations for the 3D time-dependent fluid flow, thermodynamics, and magnetic field in a rotating spherical shell that served as an analog of convection and magnetic field generation in the solar interior. The resulting solutions maintained a differential rotation at the surface, very similar to that observed on the Sun, and a magnetic field that continuously and periodically reversed as a ‘dynamo wave’ propagating in latitude, somewhat similar to the large-scale fields observed on the solar surface. However, the coarse spatial resolution, which was barely affordable at that time, forced the simulated fluid flows to be unrealistically laminar. Later it was

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**Figure 9**  An aborted reversal during the case ‘c’ simulation of Glatzmaier et al. (1999) that occurs around 88 000 years in the record of Figure 14(c). Left: True dipole path and VGP paths for five locations around the globe. Right: Longitudinally averaged poloidal flux in the outer core at the midpoint of the directional excursion shows that while the field reversed in most of the outer core and above, it retained the original normal polarity close to the inner core and within it. Note that each plotted Time Step represents about 100 years and 3500 numerical time steps.
found that the differential rotation below the surface predicted by these early models was not consistent with the profile inferred from helioseismology. Computers today provide much better spatial resolution, which allows the simulations to be at least weakly turbulent. This changes the style of the dynamics (from large convective cells to small convective plumes), the transport of angular momentum, and therefore the pattern of differential rotation, the generation of vorticity, and helicity, and ultimately the dynamo mechanism. However, the solar dynamo is still far from being well understood.

The original 3-D MHD geodynamo models were developed in the early 1990s. Surprisingly perhaps, the challenge then was not to produce reversals, but to stop them from occurring too frequently. These were strongly convective and rotationally dominant (high Rayleigh number, low Ekman number) dynamos that were quite unstable. In addition, the solid inner core (see Chapter 8.10) was originally approximated as an insulator (for numerical simplicity) because most people at that time felt the small inner core would have little effect on the dynamo in the outer fluid core. However, when Hollerbach and Jones (1993) showed that a finite-conducting inner core in their 2-D ‘mean-field dynamo model’ stabilizes their solution, Glatzmaier and Roberts (1995) made the inner core conducting in their 3-D MHD model and obtained a relatively stable, dipole-dominated dynamo (Figure 10) that, after a couple of magnetic diffusion times, produced an isolated dipole reversal (Figure 11). The large insulating inner core in their original (frequently reversing) model obstructed the flow; whereas their new conducting inner core provided an anchor for the field via magnetic torque, which resulted in a solution still time dependent but not continuously reversing. As mentioned above, tests with a much smaller conducting inner core (less flow obstruction) produced a stabler, much more antisymmetric magnetic field (Roberts and Glatzmaier, 2001; Coe and Glatzmaier, 2006), which suggests that the reversal frequency may have been very low in the distant past (see Chapter 8.02).

It has been more than a decade since that first spontaneous magnetic dipole reversal was found in a 3-D MHD self-consistent computer simulation of the geodynamo (Glatzmaier and Roberts, 1995). Since then several groups around the world have developed similar geodynamo models (see Chapter 8.08), some of which also produce magnetic reversals (Kida et al., 1997; Kageyama et al., 1999; Sarson and Jones, 1999; Kutzner and Christensen, 2002; Busse, 2002; Wicht, 2002; Takahashi et al., 2005; Reshetnyak and Steffen, 2005). These simulations are less laminar than the early solar dynamo simulations. However, since no one yet can afford the computing resources to produce a geodynamo simulation close to the Earth’s parameter regime, the simulations are still extremely crude approximations of the geodynamo. The various MHD geodynamo models employ different sets of parameters, boundary conditions, numerical methods, and spatial resolution (Glatzmaier, 2002); therefore, a detailed comparison is difficult. Yet many of the results have features quite similar to geomagnetic and paleomagnetic observations. Reviews have been written that qualitatively compare the results to the paleomagnetic reversal record (e.g., Dormy et al., 2000; Kono and Roberts, 2002).

Reversing MHD dynamo simulations fall loosely into three categories: (1) periodic dynamo waves, for which the field continuously reverses (Kida et al., 1997; Busse, 2002), (2) reversals possibly triggered by large-scale plume events, fluctuations in the axisymmetric meridional circulation, or changes in kinetic and magnetic energies (Sarson and Jones, 1999; Kida et al., 1997; Kageyama et al., 1999; Sarson and Jones, 1999; Kutzner and Christensen, 2002; Busse, 2002; Wicht, 2002; Takahashi et al., 2005; Reshetnyak and Steffen, 2005). These simulations are less laminar than the early solar dynamo simulations. However, since no one yet can afford the computing resources to produce a geodynamo simulation close to the Earth’s parameter regime, the simulations are still extremely crude approximations of the geodynamo. The various MHD geodynamo models employ different sets of parameters, boundary conditions, numerical methods, and spatial resolution (Glatzmaier, 2002); therefore, a detailed comparison is difficult. Yet many of the results have features quite similar to geomagnetic and paleomagnetic observations. Reviews have been written that qualitatively compare the results to the paleomagnetic reversal record (e.g., Dormy et al., 2000; Kono and Roberts, 2002).

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1999; Kageyama et al., 1999; Wicht and Olson, 2004; Reshetnyak and Steffen, 2005), and (3) those with relatively long epochs of varying durations between reversals of varying durations and with more frequent aborted reversals (Glatzmaier et al., 1999; Kutzner and Christensen, 2002; Takahashi et al., 2005). For the simulations of category (3) the reversals occur through the combined nonlinear action of many small-scale fluctuations.

The early solar dynamo models (Gilman, 1983; Glatzmaier, 1985) simulated the periodic and continuously occurring solar reversals as an alpha–omega dynamo wave. The more recent simulated reversals of Kida et al. (1997) and Busse (2002) also appear to be global-scale dynamo waves driven by laminar convection. However, as discussed above, paleomagnetic reversals do not continuously occur, that is, the epochs between reversals are long (~10 magnetic dipole diffusion times) compared with the typical duration of a reversal (~0.1–0.5 magnetic dipole diffusion time). (A geomagnetic dipole diffusion time is 20,000 years.) They are also aperiodic, that is, the lengths of these epochs are highly irregular (Figure 6). In addition, aborted reversals (or excursions) are much more frequent than full reversals (Figure 8). Single dipolar reversals among several aborted reversals have been simulated by Glatzmaier and Roberts (1995), Kageyama et al. (1999), and Sarson and Jones (1999).

It appears that the progression through the three categories of simulations is correlated with the vigor of the convection (i.e., the Rayleigh and Reynolds numbers) and the relative effect of rotation (i.e., the Ekman and Rossby numbers) (see Chapters 8.05, 8.03, 8.08, and 8.11). Assuming the rotation rate, electrical conductivity, heat flux, and dimensions of the Earth’s core, the smaller the prescribed thermal and viscous diffusivities the more turbulent the flow and the closer these nondimensional numbers approach Earth-like values and the more Earth-like the reversals appear. However, the character and frequency of the reversals also likely depend on several other model specifications: the ratio of buoyancy and Coriolis forces, the pattern of the heat flux over the CMB, and the relative size of the solid inner core.

Wicht and Olson (2004), for example, studied a weak-field, slowly rotating, large-scale convective dynamo, which produced a series of quite regular (periodic) reversals with periods long relative to the duration of a reversal but short relative to a magnetic diffusion time (Figure 12). These results are more Earth-like than the continuously reversing dynamo-wave solutions because of their relatively long, stable epochs between the reversals; however, the process is still periodic, that is, the durations of the epochs do not vary randomly as the Earth’s do (Figure 6). Reversed-polarity magnetic flux is generated when convective plumes twist the field; this reversed flux is then advected throughout the outer core by the dominant meridional circulation. However, Lorentz forces have very little effect on the flow in their simulations; that is, the solutions are nearly kinematic. The main advantage of this approach is the ease of analysis; that is, the reversal mechanism is large scale and the reversal frequency is high and constant. However, the reversal mechanism may not be very Earth-like.

Figure 11 Three snapshots of a simulated magnetic field (as in Figure 10) at 500 years before the mid-point in the dipole reversal, at the mid-point and at 500 years after the mid-point.
Kutzner and Christensen (2002) studied more strongly convective and more rotationally dominant cases. They found that convective dynamos with small Rayleigh numbers (i.e., small convective driving for a given rotation rate) tend to be stable and have a dominant axial dipole; large Rayleigh numbers produce frequently reversing dynamos and less dipolar fields. The cases that fall within a narrow region of parameter space between these two regimes have longer and more irregular times between reversals and therefore appear more Earth-like (Figure 13).

Glatzmaier et al. (1999) also chose strongly convective, rotationally dominant, cases to test the sensitivity of convective dynamos to the pattern of the heat-flux boundary condition on the CMB (Figure 14), presumably imposed by mantle convection. Their model

\[ Ra = 17 \times Ra_{\text{crit}} \]

\[ Ra = 27 \times Ra_{\text{crit}} \]

\[ Ra = 35 \times Ra_{\text{crit}} \]
Figure 14 Eight dynamo simulations with different imposed patterns of radial heat flux at the CMB. The top row shows the patterns of CMB heat flux. Solid contours represent greater heat flux out of the core relative to the mean; broken contours represent less. Case 'g' has a uniform CMB heat flux and case 'h' has a pattern based on seismic tomography, assuming lower sound speed corresponds to warmer mantle and therefore smaller heat flux out of the core. The second row shows the trajectory of the south magnetic pole of the dipole part of the field outside the core, spanning the times indicated in the plots below; the marker dots are about 100 years apart. The plots in the third and fourth rows show the south magnetic pole latitude and the magnitude of the dipole moment (in units of $10^{22}$ A m$^2$) vs time (in units of 1000 years). Reproduced from Glatzmaier GA, Coe RS, Hongre L, and Roberts PH (1999) The role of the Earth’s mantle in controlling the frequency of geomagnetic reversals. Nature 401: 885–890 with permission from Nature.
differs from all other geodynamo models by solving the equations of motion within the anelastic approximation instead of the Boussinesq approximation (see Chapters 8.05, 8.03, and 8.08). That is, in their model the variation of density with depth is taken into account and both compositional and thermal buoyancy are computed. In addition, more self-consistent thermal and compositional boundary conditions are applied at the inner-core boundary. They found that forcing greater heat flux through the CMB in the polar and equatorial regions, opposed to mid-latitude, produces a strong, stable, axial dipole, whereas the opposite CMB heat flux pattern produces frequent reversals (see Chapter 8.12). Of the eight cases they tested, their Earth-like tomographic heat flux case (h) and their homogeneous heat flux case (g) appear most Earth-like in terms of the long and irregular times between reversals and the relatively short and highly variable reversal durations. In addition, like the Earth, case ‘h’ has more frequent aborted reversals than successful reversals. The system seems to be continually trying to reverse via MHD instabilities but only after many attempts are the conditions favorable for the new polarity to continue to grow and the old polarity to be fully destroyed (Figure 15). This continuing process of trying to reverse occurs in these simulations at roughly the frequency at which aborted reversals occur in the Earth, which is similar to the frequency that regular full reversals occur in the Wicht and Olson (2004) simulations (Figure 12). However, unlike the reversals seen by Wicht and Olson (2004), many of these simulated reversals differ in terms of morphology and duration.

As computational resources continue to improve, simulations using much smaller (i.e., more realistic) diffusivities become possible. Takahashi et al. (2005) produced some of the most realistic geodynamo simulations to date by using much larger Rayleigh numbers and smaller Ekman numbers, that is, smaller, more realistic viscous, thermal, and magnetic diffusivities. Several dipole reversals occurred with a highly variable reversal frequency (Figure 16). They also show that the internal dynamics of the fluid core changes significantly when the viscous forces become much smaller than the Coriolis (rotational) and Lorentz (magnetic) forces, which underscores the importance of reaching the very low-viscosity (turbulent) regime if one wishes to simulate and understand the dynamo mechanism in the Earth’s core.

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![Magnetic field reversal](image)

**Figure 15** A sequence of snapshots of the longitudinally averaged magnetic field through the interior of the core and of the radial component of the field at the CMB and at what would be the surface of the Earth, displayed at roughly 3000-year intervals, spanning the first dipole reversal of case ‘h’ in Figure 14. In the plots of the average field, the small circle represents the inner-core boundary and the large circle is the CMB. The poloidal field is shown as magnetic field lines on the left-hand sides of these plots (blue is clockwise and red is counter-clockwise). The toroidal field direction and intensity are represented as contours (not magnetic field lines) on the right-hand sides (red is eastward and blue is westward). Hammer (equal area) projections of the entire CMB and surface are used to display the radial field (with the two different surfaces displayed as the same size). Reds represent outward-directed field and blues inward field. The surface field, which is typically an order of magnitude weaker, was multiplied by 10 to enhance the color contrast. Adapted from Glatzmaier GA, Coe RS, Hongre L, and Roberts PH (1999) The role of the Earth’s mantle in controlling the frequency of geomagnetic reversals. *Nature* 401: 885–890.
8.09.3 Conclusions

Our understanding of geomagnetic reversals has improved considerably over the years with paleomagnetic studies and geodynamo simulations. Paleomagnetic observations now provide considerable constraints on the time-averaged field and the character of reversals, some of which have been matched to first-order in some dynamo simulations. Nonetheless, more ‘ground truthing’ observations and more realistic simulations are needed to discover the details of the reversal mechanism in the Earth’s core and what influences its range of variation.

In terms of observations, longer, more reliable, and more detailed records of paleomagnetic field behavior are needed, especially early in Earth’s history. This of course will take considerable time and effort.

In terms of models, geodynamo simulations need to be more turbulent and rotationally dominant. Laminar flows are certainly easier to produce and analyze; but it is unlikely that the large-scale convection cells seen in such studies produce a reversal mechanism representative of that in the Earth’s turbulent outer core. The same conclusion has been reached in the solar dynamo community. In addition, much longer simulations are needed to gather better statistics on the types and frequencies of reversals. Finally, the sensitivities to several model specifications, like density stratification, boundary conditions, compositional buoyancy, inner-core conductivity, and diffusion coefficients, need to be tested. These modeling goals all require significant amounts of computing resources and, as in the past, compromises will need to be made as we wait for computer hardware to improve. Support for this research was provided by NSF CSEDIP program.

References


