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8.11 Magnetic Polarity Reversals in the Core

8.11.1 Introduction

Paleomagnetic observations have provided much information about the nature of geomagnetic dipole reversals and three-dimensional (3D) magnetohydrodynamic (MHD) computer models are beginning to provide insight to the reversal mechanism. Reversals typically take 2000–12000 years to occur—one to two orders of magnitude shorter than the dipole-dominated, stable polarity epochs between reversals. However, the durations of both the polarity transitions and epochs vary greatly. The average reversal frequency appears to depend on the degree of equatorial symmetry of the nonaxial dipole (NAD) field (i.e., the total field minus the axial dipole term) and may have been greatly reduced prior to ~1 Ga. Virtually all records document lowered intensity during reversals, but they differ in complexity of transitional direction change. At least some reversals have highly complex transition paths. In addition, stable polarity epochs are often punctuated by large directional excursions and reduced intensity that occur more frequently than reversals and may be regarded as failed reversals. A statistical bias of transition poles for preferred geographical regions that correlate with lower-mantle tomography has been claimed, and such bias has been produced in dynamo simulations employing laterally varying heat flux at the core–mantle boundary (CMB). Higher-fidelity records at multiple locations over the globe are needed to establish how the field varies in time and space during reversals and to provide guidance for computer modeling. No external trigger is necessary for reversals, which occur as a natural result of the very nonlinear dynamo process; that is, the flow, field, and thermodynamic variables of the Earth’s outer fluid core all depend on time and space and on each other in a chaotic way. The more realistic computer simulations suggest that MHD instabilities continually occur but only one in many attempts results in a complete reversal of the magnetic dipole field and that these successful reversals often have different morphologies and durations. Computer models have provided new insights into the mechanism of the geodynamo and its reversals, but much better spatial resolution, lower viscosity, and longer simulations are needed to reveal and understand the complex turbulent processes that underlie the polarity reversal phenomenon.

8.11.2 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 5.01–5.07; see also Chapters 5.08–5.14). Thus, for information about most geomagnetic behavior, 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 My 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 Precambrian, but Precambrian evidence is sparse and equivocal (Dunlop and Yu, 2004). To the 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 Ma, the departure can be represented as an axial quadrupole contribution that is 3–5% of, and the same sign as, the axial dipole (Johnson et al., 2008; McElhinny, 2004). The 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, Matuyama–Brunhes reversal, which occurred 0.78 Ma, the error in age for the best-dated transitional basalt flows is comparable to the transition duration (Singer et al., 2005). For this reason, the duration of reversals 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 transitional field has a large nondipolar component.

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 sediment records show that at least some reversals are complex, with episodes of oscillatory and rapid field change (Figures 2(a) 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 the estimation of duration. For instance, if one includes an early unsuccessful swing to normal polarity in the Matuyama–Brunhes 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(a) and 3. Even excluding the precursor, high-resolution sediment records show that this transition is complex, with multiple large swings of the field direction (Channell and Lehman, 1997). Some dynamo simulations have also produced long, complex 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).

Despite the striking differences in high-resolution reversal records, a recent study by Valet et al. (2012) boldly advances the hypothesis that all polarity transitions are governed by a common underlying process in which the dipole field collapses and regenerates three times in close succession (Figure 5(a) and 5(b)). During each phase of low dipole intensity, the remaining nondipole field produces highly variable directions between normal and reversed polarity. The first phase is a precursor in which the dipole regenerates in its original polarity, the second phase is the transit to opposite polarity, and the third phase is a rebound in which the dipole decays and regenerates in its recently achieved opposite polarity. In support of their conceptual model, they consider only the most detailed lava-flow transition zones, thereby avoiding possible problems of smoothing and other artifacts of magnetic recording in sedimentary records, and require that they occur in stratigraphically ordered volcanic sequences with full-polarity normal and reversed flows above and below. Furthermore, they assume that the average interval between successive flows for each record, though quite different from one record to another, is more or less constant within a single record. This allows them to stack the records as shown in Figure 5(a), rescaling the lava-flow numbers to a common correlation number by lining up the records at the first transitional flow, stretching or compressing each of them so that they also line up at the final transitional flow or at some other distinctive point in the records, and plotting the lava-flow correlation number versus reversal angle for each record. Observations and modeling of recent field variability from archeomagnetic records lead to estimates of ~2.5 ky or less for the precursor and rebound and ~1 ky for the transit. Their estimate of 5–6 ky for the duration of an entire reversal accords well with most estimates, including the average duration of 7 ky of Clement (2004), but of course not with the 2–12 ky latitudinal variability (see Figure 1). It also fits with the durations of reversals exhibited in many dynamo simulations (e.g., Glatzmaier et al., 1999; Kutzner and Christiansen, 2004; Olson et al., 2011), though there are significant exceptions (e.g., simulation (h) of Figure 14 where the second reversal takes four to five times longer). Six of the seven records in Figure 5(a) exhibit some degree of directional rebound (reversal angle in excess of 30°), whereas only three records show a precursor. However, one of those shown in Figure 5(a) without a precursor (Steens Mountain) is not an exception because there are actually three successive lava flows a little below the transit zone that do have paleomagnetic reversal angles slightly >30° (Mankinen et al., 1985; Table 1 lavas B70–72). Moreover, in the lava-flow section slightly below the transit zone in the Hawaii 0.78 My record, radiometric dating has revealed an ~100 ky hiatus (Coe et al., 2004), which can explain the discrepancy with the Tahiti 0.78 My record that does display a precursory mode.

Despite its apparent success, how generally applicable this three-phase model is to reversals will require further testing. Dynamo simulations have produced numerous exceptions to the rule, with transition paths that are too simple or too complex, but what is needed are more high-resolution...
paleomagnetic reversal records of the full-field vector, intensity, and direction. The most detailed one available currently is the Steens volcanic record, with an apparent resolution two or more times that of the other detailed volcanic records (Figure 5(c)) and considerable paleointensity data (Prévot et al., 1985). It exhibits the additional complexity of two closely spaced directional rebounds rather than only one (Figure 2), but intensity data are lacking to show whether or not the field recovered its strength during the interval between the directional rebounds.

In the search for some kind of systematics to the tangle of reversal transition paths, various authors have long proposed that there is a tendency for transitional virtual geomagnetic poles (VGPs) to cluster geographically in preferred longitudinal bands or patches on Australasia and the Americas and to avoid the Central Pacific Basin (Clement, 1991; Hoffman, 1992; Hoffman and Mochizuki, 2012; Hoffman and Singer, 2004; Laj et al., 1991; Love, 2000), though the statistical significance of this tendency has been disputed (McFadden et al., 1993; Merrill and McFadden, 1999; Prévot and Camps, 1985).

Figure 2  (a) Composite directional record of the 16.7 Ma Steens Mountain reversal recorded in four sections of superposed lava flows in SE Oregon (Jarboe et al., 2011). Arrows indicate the progression of time, and open (solid) symbols indicate upward (downward)-pointing directions. Note the increased complexity of the transition path when new directions found at Poker Jim Ridge (north and south) and Catlow Peak are spliced into the earlier record from Steens Mountain of Mankinen et al. (1985) and Camps et al. (1999). Open and closed stars are the reversed and normal axial dipole directions for that region. (b) The transition path of virtual geomagnetic poles (VGPs) for the Steens Mountain reversal derived from the composite directional record of Figure 2(a) (Jarboe et al., 2011). Some of the stop-and-go character of much of the path, with clusters of VGPs separated from each other by sizable gaps, probably reflects the episodic nature of volcanic eruptive activity, but the clusters around the west coast of South America and in the northwestern Pacific Ocean may represent significant lengths of time (see text).
If true, the long timescale would suggest the influence of the lowermost mantle. The preferred areas do correlate with large-scale seismic tomography, overlying regions in the lower mantle of higher than average P- and S-wave velocity (e.g., Grand, 2002). 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 been produced by some dynamo simulations employing appropriate heat-flux boundary conditions at the CMB (Coe et al., 2000; Kutzner and Christensen, 2004; Olson and Christensen, 2002).

The Steens reversal record (Figure 2(b)) does not exhibit a longitudinal band or bands, but rather a series of clusters of VGPs, raising the question whether the clusters record standstills in the field or episodic volcanism. It is well known that eruptive activity is generally episodic, but the two most prominent VGP clusters of the Steens record, in western South America and the western Pacific, are each represented by multiple flows from sections 60 to 80 km apart. Moreover, the chemistry of the flows differs significantly both within and between sections, and the South American cluster is visited twice, both early and late in the record. Rather than brief bursts of volcanism, lingering of the VGP for appreciable periods of time seems a better explanation for these clusters. This in turn suggests geographic preference, which is consistent with the higher than average S-wave velocities detected beneath them in the lowermost mantle (Grand, 2002).

There are clusters of VGPs at these locations in some other polarity transitions (see Jarboe et al., 2011 for examples), as would be expected if clustering is due to the effect of low-temperature anomalies in the lowermost mantle, but that raises the question why there are VGP paths for plenty of other polarity transitions that do not display clustering in these regions (e.g., see Figure 1 in Valet et al., 2012). Some of these transitions do have VGP clusters in other places (e.g. Glen et al., 1999; Hoffman and Mochizuki, 2012; Jarboe et al., 2011) such as around Australia and eastern North America, both of which are characterized by fast lower-mantle shear-wave anomalies, but sometimes also in northern Africa, which paradoxically is underlain by slow shear-wave velocities (Grand, 2002). Support for transitional VGP clustering is also sought in the vertical component of the modern NAD field – that remaining after

![Figure 3](Image)

*Figure 3* The Gauss–Matuyama (2.58 Ma) reversal record of VGPs recorded in sediments deposited in Searles Lake, California (Glen et al., 1999). 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](Image)

*Figure 4* 40Ar/39Ar 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 those of 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.

If true, the long timescale would suggest the influence of the lowermost mantle. The preferred areas do correlate with large-scale seismic tomography, overlying regions in the lower mantle of higher than average P- and S-wave velocity (e.g., Grand, 2002). 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 been produced by some dynamo simulations employing appropriate heat-flux boundary conditions at the CMB (Coe et al., 2000; Kutzner and Christensen, 2004; Olson and Christensen, 2002).

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Figure 5  Sequential lava-flow numbers for the most complete, stratigraphically ordered lava-flow reverse-to-normal polarity transition records are rescaled to a common correlation number by assuming that the polarity transitions have a common duration. In (a), the correlation number is plotted versus reversal angle for these records and suggests that an underlying structure involving three phases of field instability might be common to reversal transitions, as shown in (b) and discussed in the text. (c) The degree of detail of the reversal records is indicated by the number of distinct field directions recorded by groups of sequential lava flows in the transition zone. Adapted from Valet JP, Fournier A, Courtillot V, and Herrero-Bervera E (2012) Dynamical similarity of geomagnetic field reversals. Nature 90: 89–94.
subtraction of the axial dipole field, which exhibits strong downward-directed anomalous patches in eastern North America and the southern Atlantic Ocean and an upward-directed patch in the Australasian region (Hoffman and Mochizuki, 2012; Figure 4). These flux patches cause clustering of VGPs of the modern NAD field in eastern North America and the southern Atlantic regions and also in the region antipodal to Australasia. Under the assumption that these modern flux patches are caused by lower-mantle anomalies, as proposed by Hoffman and Singer (2008), then they would be expected to persist for tens of millions of years and produce VGP clustering of the paleomagnetic transitional field in these regions or their antipodes, depending on whether the patches contained upward or downward flux. A problem is that these six regions where one could expect flux patches under this scenario occupy a very appreciable proportion of the total area in which a VGP would be classified as transitional. Thus, the degree to which transitional VGP clusters demonstrate statistically significant geographic preference caused by lower-mantle anomalies remains difficult to assess rigorously.

Earth’s reversals occur aperiodically, in fact almost randomly, but the mean duration appears to change progressively over long time intervals (Figure 6(a)). As determined from dating rocks of known magnetic polarity and from analyzing the record of marine magnetic anomalies produced by sea-floor spreading, from 0 to 165 Ma, the field reversed on average about two times per million years, but over 10 My intervals the average reversal rate varied from highs of at least five per million years 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. Two other superchrons have been documented in the past 550 My, the Kiaman Reversed and the Moyero Reversed superchrons (Pavlov and Gallet, 2005) in Permo-Carboniferous and Ordovician time, respectively (Figure 6(b)).

Figure 6 (a) Smoothed reversal rate (thin wavy line) corresponding to normal and reversed chron (black and white bars) of the geomagnetic polarity timescale (Ogg et al., 2004). The rate goes to zero in the Cretaceous Normal Superchron, from 83 to 118 Ma. The squares denote \( b/a \), a relative measure of the average ratio of the antisymmetric to symmetrical 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 (\( \sim 10 \) My\(^{-1} \)) (adapted from Coe RS and Glatzmaier GA (2006) Symmetry and stability of the geomagnetic field. Geophysical Research Letters 33: L21311. http://dx.doi.org/10.1029/2006GL027903). (b) In the last 550 My, there are two well-documented superchrons when no true polarity reversals have been unequivocally documented and a third, the Moyero Reversed Superchron, that has been proposed more recently (adapted from Pavlov V, Gallet Y (2005) A third superchron during the Early Paleozoic. Episodes 28: 1–7).
From the 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 NAD 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(a)). For example, the equatorial dipolar and axial quadrupolar harmonics of the ﬁeld are symmetrical and the axial octupolar harmonic is antisymmetric (Merrill, 1996). The analysis of the Glatzmaier et al. (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 symmetrical gauss coefﬁcients (Coe and Glatzmaier, 2006). More recently, a new analysis of the latitude dependence of secular variation for an expanded set of lava ﬂows erupted during the CNS reafﬁrms this correlation between lower reversal rate and higher antisymmetric harmonic content of the ﬁeld (Biggin et al., 2008a). In addition, an insightful synthesis of new and previous studies of Permo-Carboniferous red beds came to a similar conclusion for the Kiaman Reversed Superchron (Haldan et al., 2009).

Relatively little is known for sure about the ancient, Precambrian geomagnetic ﬁeld, but there is evidence indicating that it existed as early as 3.40–3.45 Ga (Tarduno et al., 2010), and the earliest documented reversal occurred around 3.2 Ga (Layer et al., 1998). Several attempts have been made to apply the same secular variation approach applied in the preceding text to make inferences about reversal frequency in the Precambrian. A dynamo simulation with a solid inner core that is only one-quarter of its size today (Roberts and Glatzmaier, 2001) produced an even more antisymmetric NAD ﬁeld (Coe and Glatzmaier, 2006), hinting that reversals may have been much less common in the distant geologic past when the Earth’s inner core was smaller (see Chapter 8.02). Limited paleomagnetic evidence available documenting reversals in several thick sections of rocks older than 1 Ga and less than 2.8 Ga suggested an average reversal rate ﬁve to ten times less than the well-documented rate of ~1.7 per million years for the 0–150 Ma interval (Coe and Glatzmaier, 2006, and references therein), although large temporal gaps between these sections and likely large hiatuses could make this estimate unrepresentative. But two later studies of Precambrian secular variation in lavas and dikes between 1.1 and 2.8 Ga provide additional indirect support for a slower Precambrian reversal rate by inferring a greater degree of antisymmetry of the average ﬁeld relative to the modern ﬁeld (Biggin et al., 2008b; Smirnov et al., 2011).

Most recently, on the basis of their magnetostratigraphic ﬁndings from Siberia and those of Elston et al. (2002) from North America, Gallet et al. (2012) suggested that the ﬁeld before ~1 Ga was characterized by a greater proportion of superchrons that are bounded abruptly by periods of high reversal frequency, but the average reversal rate estimated from 1 to 2.8 Ga is nonetheless much lower than from 0 to 150 Ma. Thus, the emerging consensus (see Nimmo vol. 8 Ennergetics of the Core and vol. 9 Thermal and Compositional Evolution of the Core) that inner-core nucleation occurred relatively recently in earth history, around 1 Ga or even later, is consistent with the observations suggesting signiﬁcantly different, on-average lower reversal frequency during Mesoproterozoic and Neoarchean time. Although these observations are uncertain, the computer simulations of the geodynamo mentioned earlier in the text support a lower reversal frequency with a smaller solid inner core. The main reason for this model result is the deep-shell geometry of the ﬂuid outer core: that is, a smaller inner core provides less obstruction within the ﬂuid core and therefore weaker instabilities in the global ﬂow and ﬁeld (see Section 8.11.3).

In addition to reversals, which entail a switch from one well-established polarity state to stable opposite polarity, the geomagnetic ﬁeld has undergone many brief excursions to reduced intensity and reversed or nearly reversed directions, some of which have been detected around the globe. These might represent failed reversal attempts. During the past 0.78 My of today’s normal polarity epoch when global simultaneity of brief ﬁeld ﬂuctuations is easier to establish, there have been eight or more excursions (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 of ~1–10,000 years. In the paleomagnetic and sea-ﬂoor magnetic anomaly records, their classiﬁcation is somewhat arbitrary (Acton et al., 2006): It is diﬁcult to draw a line between failed reversals and very short true reversals on the one hand and

![Energy Distribution](https://dx.doi.org/10.1029/2006GL027903)

**Figure 7** Time-averaged spatial energy density $W_{nm}^m$ 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 symmetrical 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. http://dx.doi.org/10.1029/2006GL027903.
unusually large directional secular variation on the other. However, in dynamo simulations, failed reversals can be unambiguously defined as excursions 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, as illustrated in Figure 9 (Glatzmaier et al., 1999).

### 8.11.3 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

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**Figure 8** Excursions of the geomagnetic field since the last reversal (Brunhes–Matuyama), as indicated by deep minima in the 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.

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**Figure 9** A failed reversal during case ‘c’ simulation of Glatzmaier et al. (1999) that occurred 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.
thermodynamics and 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 2D axisymmetric models of the mean magnetic fields. These models are kinematic, that is, the fluid flow is prescribed or parameterized instead of being part of the solution. Only the longitudinally averaged part of 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 parameterizes 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, 3D, time- and spatially dependent processes with 2D 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 3D dynamics must be that maintains the prescribed alpha and omega effects. Discovering and understanding the details of the dynamo mechanisms require a self-consistent solution of the full set of nonlinear 3D 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 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 3D 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 and 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 2D ‘mean-field dynamo model’ stabilizes their solution, Glatzmaier and Roberts (1995) made the inner core conducting in their 3D 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

Figure 10 A snapshot of the simulated geomagnetic field produced by Glatzmaier and Roberts (1995). A set of magnetic lines of force illustrated the 3D structure of the field, which is intense and complicated inside the fluid core and smooth and dipole-dominated outside the core. The rotation axis of the model Earth is vertical in the illustration and yellow lines represent outward-directed field and blue lines represent inward-directed field. The field is sheared around the ‘tangent cylinder’ to the inner-core equator.
mentioned in the preceding text, tests with a much smaller conducting inner core (less flow obstruction) produced a more stable and more antisymmetric magnetic field (Coe and Glatzmaier, 2006; Roberts and Glatzmaier, 2001), which suggests that the reversal frequency may have been very low in the distant past (see Chapter 8.02).

It has been nearly two decades since that first spontaneous magnetic dipole reversal was found in a 3D 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 (Busse, 2002; Kageyama et al., 1999; Kida et al., 1997; Kutzner and Christensen, 2002; Reshetnyak and Steffen, 2005; Sarson and Jones, 1999; Takahashi et al., 2005; Wicht, 2002). 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 (Busse, 2002; Kida et al., 1997); (2) reversals possibly triggered by large-scale plume events, fluctuations in the axisymmetric meridional circulation, or changes in kinetic and magnetic energies (Kageyama et al., 1999; Reshetnyak and Steffen, 2005; Sarson and Jones, 1999; Wicht and Olson, 2004); and (3) those with relatively long epochs of varying durations between reversals of varying durations and with more frequent failed 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) produced periodic and continuously occurring solar reversals via an alpha–omega dynamo-wave mechanism. The more recently 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 in the preceding text, 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 20000 years.) They are also aperiodic; that is, the lengths of these epochs are highly irregular (Figure 6). In addition, failed reversals (or excursions) are much more frequent than full reversals (Figure 8). Several of the early MHD dynamo simulations produced qualitatively similar sequences of full and failed dipole reversals (Glatzmaier et al., 1999; Kageyama et al., 1999; 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 8.03, 8.05, and 8.08). 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

Figure 11 Three snapshots of a simulated magnetic field (as in Figure 10) at 500 years before the midpoint in the dipole reversal, at the midpoint, and at 500 years after the midpoint.
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.

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). Increased reversal frequency with increased convective driving relative to the effects of rotation seems quite reasonable. Polarity reversals in dynamo simulations tend to be initiated by chaotic modifications in the dominant fluid flow pattern, like the differential rotation, that is maintained by Coriolis forces organizing the structure of convectively driven flow. These temporary flow modifications twist and shear the existing magnetic field in ways that are slightly different than the usual dynamo mechanism; this destroys the original polarity and begins to generate the new polarity. Such perturbations in the flow need to continue long enough for the original polarity to be completely destroyed and replaced by a global field with the new polarity. The tendency for such modifications in the flow is greater when the chaotic effects of the convection are greater than the organized effects of the rotation.

| Ra = 17 \cdot Ra_{crit} |
| Ra = 27 \cdot Ra_{crit} |
| Ra = 35 \cdot Ra_{crit} |

Figure 12 The magnetic pole latitude (at the surface) versus time (in magnetic diffusion times, here estimated as 80,000 years). Adapted from Wicht J and Olson P (2004) A detailed study of the polarity reversal mechanism in a numerical dynamo model. Geochemistry, Geophysics, Geosystems 5. http://dx.doi.org/10.1029/2003GC000602.

Figure 13 Three time series of a simulated field. The top row is for a low Rayleigh number, which is stable. The middle row is for a higher Rayleigh number, which has failed reversals. The bottom row is for a still higher Rayleigh number, which shows dipole reversals and failed reversals. Each row shows the dipole colatitude at the surface versus time (in magnetic diffusion times, here estimated as 80,000 years). Adapted from Kutzner C and Christensen U (2002) From stable dipolar to reversing numerical dynamos. Physics of the Earth and Planetary Interiors 121: 29–45.
Several other groups have since studied this reversal-tendency dependence on the relative effects of convection and rotation by analyzing many long dynamo simulations (e.g., Christensen and Aubert, 2006; Drixell and Olson, 2009; Lhuiller et al., 2013; Olson and Christensen, 2006; Olson et al., 2010, 2011). They all basically agree that in parameter space, a transitional regime exists between one dominated by rotation, characterized by a dominantly dipolar field structure (above the CMB) that does not reverse, and one dominated by convection, characterized by a multipolar field structure with a frequently reversing dipole. Dynamo simulations in this transitional parameter regime have long stable dipole-dominated chrons separated by relatively short reversals, which sometimes are very complex involving failed reversals (Lhuiller et al., 2013; Olson et al., 2011). These studies suggest that the Earth’s dynamo sits in an extension of this transitional regime where the effects of convection and rotation are comparable but both are many orders of magnitude greater than what they are for any of these computer-simulated dynamos. Although quite possible, such a huge extrapolation in parameter space is necessarily uncertain because the characteristics of a dynamo mechanism depend not only on the ratio of the convective to rotational effects but also on the absolute amplitudes of these two fundamental dynamo ingredients.

Glatzmaier et al. (1999) 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 differs from all other geodynamo models by solving the equations of motion within the anelastic approximation instead of the Boussinesq approximation (see Chapters 8.03, 8.05, 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, as opposed to midlatitude, 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 failed 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 failed 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). As mentioned earlier in the text, Coe and Glatzmaier (2006) (Figure 7) show that, for the cases tested by Glatzmaier et al. (1999), low reversal and excursion frequencies and high field intensities are associated with magnetic fields at the CMB that are mainly antisymmetric with respect to the equator (i.e., more like an axial dipole field instead of an axial quadrupole field).

Several other dynamo simulation studies with similar patterns of heterogeneous CMB heat flux have since been made; these span many more magnetic diffusion times and produce more reversals per case than does the Glatzmaier et al. (1999) study. However, some of the conclusions from these later studies conflict with those of Glatzmaier et al. (1999). For example, Kutzner and Christensen (2004) and Olson et al. (2010) found that reducing the CMB heat flux in the equatorial region reduces the resulting reversal and excursion frequencies in their dynamo simulations and found that their reversal frequencies are fairly insensitive to the relative amount of CMB heat flux in the polar regions or to the nonaxissymmetry of the heat-flux boundary condition. In particular, the most stable (i.e., nonreversing) case of Glatzmaier et al. (1999) is their case E (Figure 14(e)), which has an axissymmetric CMB heat flux peaking in both the equatorial and polar regions; this case also maintains the highest field intensity, dipolarity, and axial antisymmetry (Table 1 of Coe and Glatzmaier, 2006). In contrast, this CMB heat-flux pattern produces relatively high reversal and excursion frequencies in the simulations by Kutzner and Christensen (2004) and by Olson et al. (2010). The three studies do, however, agree that a CMB heat-flux pattern like that of Case D in Figure 14(d) (i.e., high CMB heat flux in the polar regions and low flux in the equatorial region) produces a lower reversal frequency than the one like that of Case C, which has the opposite CMB heat-flux pattern of Case D imposed. This may suggest that the region of parameter space sampled by Glatzmaier et al. (1999) is more sensitive to the heat flux in the polar regions, whereas that sampled by Kutzner and Christensen (2004) and Olson et al. (2010) is more sensitive to the heat flux in the equatorial region. Kutzner and Christensen (2004) and Olson et al. (2010) also investigated what effect the prescribed amplitude of the spatial variations in the CMB heat flux has on the reversal frequency in their simulations; both studies found little effect.

Another issue, for which various simulation studies have produced conflicting conclusions, is the effect of a finitely conducting solid inner core on the frequency of dipole reversals. As mentioned in the preceding text, Glatzmaier and Roberts (1995) found that an insulating inner core in their early simulations produced a reversal frequency much greater than that seen in the paleomagnetic reversal record and that making their inner core electrically conducting reduced their reversal frequency. The long magnetic diffusion time of the inner core (a couple thousand years), relative to the shorter convective turnover time of the outer core (a couple hundred years), tends to anchor the original polarity and increase the time fluid flow modifications in the outer core need to persist in order to achieve a successful reversal. That is, like the stabilizing effects of rotation mentioned previously in the text, the conducting inner core inhibits changes in the global magnetic field structure that convection attempts to cause. However, two subsequent dynamo simulation studies produced conclusions about the effects of a conducting inner core on the frequency of magnetic reversals that differ from those of Glatzmaier and Roberts (1995) and Glatzmaier et al. (1999). Wicht (2002) found very little difference between his insulating inner-core case and his conducting inner-core cases. Lhuiller et al. (2013), on the other hand, found that the conductivity of the inner core does have a significant effect, but the opposite of that found by Glatzmaier and Roberts (1995); that
is, Lhuiller et al. (2013) found that their insulating inner-core cases have a smaller reversal frequency than their conducting inner-core cases. Lhuiller et al. (2013) provided a detailed statistical analysis, based on empirical probability distribution functions of their simulated results, that illustrates how their insulating inner-core case is much more likely to be in one of the two stable axial dipole polarity states than in an unstable transition state. Their conducting inner-core cases, on the other hand, show a higher probability of being in the transition state, that is, they have higher reversal and failed reversal frequencies.

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 that 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; 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 $10^{22}$ Am$^2$) versus 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.
A related issue is how the size of a conducting solid inner core affects the reversal frequency. Dynamo simulations by Roberts and Glatzmaier (2001) that investigate the past, present, and future of the geodynamo suggest that the Earth’s magnetic field in the distant past, when the solid inner core was smaller or did not yet exist, may have been more strongly dominated by the axial dipole and less likely to experience dipole reversals than it is today; this trend is suggested to continue into the distant future. The reason for this is that a smaller inner core provides less obstruction between hemispheres for fluid flow and magnetic field, resulting in weaker perturbations to the global flow and field. As discussed in Roberts and Glatzmaier (2001) that investigate the past, prehistoric, and future of the geodynamo suggest that the Earth’s magnetic field in the distant past, when the solid inner core was smaller or did not yet exist, may have been more strongly dominated by the axial dipole and less likely to experience dipole reversals than it is today; this trend is suggested to continue into the distant future. The reason for this is that a smaller inner core provides less obstruction between hemispheres for fluid flow and magnetic field, resulting in weaker perturbations to the global flow and field. As discussed in

Roberts and Glatzmaier (2001) that investigate the past, present, and future of the geodynamo suggest that the Earth’s magnetic field in the distant past, when the solid inner core was smaller or did not yet exist, may have been more strongly dominated by the axial dipole and less likely to experience dipole reversals than it is today; this trend is suggested to continue into the distant future. The reason for this is that a smaller inner core provides less obstruction between hemispheres for fluid flow and magnetic field, resulting in weaker perturbations to the global flow and field. As discussed in

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 counterclockwise). 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-directed 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.

Figure 16 The magnetic pole latitude (at the surface) versus time (in magnetic diffusion times, here estimated as 200 000 years). Adapted from Takahashi F, Matsushima M, and Honkura Y (2005) Simulations of a quasi-Taylor state geomagnetic field including polarity reversals on the Earth Simulator. Science 309: 459–461.
partly due to the differences in dynamo model approximations. Some assume a constant background density; some account for density stratification. Some simulate only thermal convection or only compositional convection; some define a codensity that approximates the combination of the two; some treat both separately, each making its own contribution to buoyancy and to the flux at the inner-core and outer-core boundaries. Some treat the solid inner core as an insulator; some treat it as a conductor, but may or may not allow it to freely rotate in response to the viscous and magnetic torques on it.

However, the main reason for the disagreement among simulation studies is likely due to the approximations those designing the dynamo simulations are forced to make because parallel supercomputers are not yet fast enough to simulate global MHD convective dynamos in a parameter regime that would closely approximate the strong rotationally dominant turbulence that exists in the Earth’s low-viscosity fluid outer core. That is, all ‘geodynamo’ simulations to date have some control parameters, like viscosity, set to values far from what would produce realistic flows and fields. The various studies choose different ways to do this. For example, most studies choose viscous, thermal, compositional, and magnetic diffusivities that are all roughly equal to the magnetic diffusivity of the Earth’s core, instead of setting the viscous, thermal, and compositional diffusivities to more realistic values orders of magnitude smaller. They define ‘time’ based on the correct (20,000-year) magnetic diffusion time, which is proportional to the square of the radius of the Earth’s core divided by its magnetic diffusivity. However, the ratio of viscous to rotational effects, that is, the Ekman number, in all dynamo simulations, especially those used to analyze long simulated times required for reversal studies, is forced (for computational reasons) to be many orders of magnitude larger than it is for the Earth’s fluid core. Therefore, the rotational period in these models is four to five orders of magnitude longer than the Earth’s rotational period of 1 day. An alternative (Glatzmaier and Roberts, 1995; Glatzmaier et al., 1999) is to use not only the Earth’s correct magnetic diffusion time but also its actual 1-day rotation period. To do this (with current computers) requires using a smaller Ekman number and a viscous diffusivity much larger than the Earth’s magnetic diffusivity (instead of smaller than it). The smaller Ekman number requires a much smaller numerical time step (because of an inertial oscillation constraint proportional to the rotation period of the model core) and a larger Rayleigh number, which requires much greater spatial resolution. This, in turn, significantly increases the computational resources required to simulate a magnetic diffusion time; but ‘time’ then is geophysically realistic for the two major forces in the problem: Coriolis (i.e., rotational) and Lorentz (i.e., magnetic). Obviously, neither way of coping with the lack of sufficient computational power should be expected to produce realistic geodynamo simulations (Glatzmaier, 2002); and it is not surprising that these different choices result in somewhat different conclusions about what controls the reversal frequency of the geodynamo.

A goal for future dynamo simulations (Glatzmaier, 2002) is to be able to run with a viscous diffusivity no greater than the magnetic diffusivity of the Earth’s core (2 m$^2$s$^{-1}$) and at the rotation period of 1 day, which would mean an Ekman number of 10$^{-9}$, about three orders of magnitude smaller than what can be done today in a global 3D dynamo simulation. Efforts are currently being made to improve or completely rewrite dynamo codes to take advantage of future supercomputers on which hundreds of thousands of processors will be able to work efficiently in parallel on a dynamo simulation that runs for months at a time simulating the geodynamo at a high enough spatial resolution and temporal resolution to capture strong turbulence and for a long enough simulated time to study reversals.

### 8.11.4 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 the 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 and codes to improve. Support for this research was provided by NSF CSEDI and Geophysics programs.

### References


