Modeling the Earth’s Dynamo

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For the past decade, three-dimensional time-dependent computer models have been used to predict and explain how the geomagnetic field is maintained by convection in the Earth’s fluid core. Geodynamo models have simulated magnetic fields that have surface structure and time dependence similar to the Earth’s field, including dipole reversals. However, no dynamo model has yet been run at the spatial resolution required to simulate a broad spectrum of turbulence, which surely exists in the Earth’s fluid core. Two-dimensional simulations of magnetoconvection show how the structure and time dependence of even the large-scale features change dramatically when the solution becomes strongly turbulent. Although these two-dimensional turbulent simulations lack the important effects of three-dimensional spherical geometry, based on their results one must question how geophysically realistic the large-scale dynamo mechanism is in current three-dimensional laminar simulations. Whatever the answer, we look forward to new discoveries from the next generation of turbulent dynamo models.

1. BASIC GEODYNAMO THEORY

We spend our entire lives immersed within the Earth’s magnetic field yet most people today seldom stop to ponder about what it is, why it exists, how it changes or how life would be different without it. The geomagnetic field helps to shield us from cosmic radiation and for ages has been used for navigation. However, the main reason for studying the geodynamo is to gain a better understanding and appreciation for the origin and evolution of this fascinating field we live in.

It has long been known that the geomagnetic field originates in the Earth’s core. However, the temperature of the core is well above the Curie temperature for permanent magnetism; the field would decay away with a half-life of about 15,000 years if it were not continually being regenerated. The Earth’s magnetic field must therefore be maintained by large electric currents within its iron-rich fluid core.

The Earth is cooling and the resulting drop in temperature with radius through the fluid core is steep enough to drive thermal convection. In addition, as the core cools, iron in the iron-alloy fluid preferentially plates onto the solid inner core releasing latent heat and leaving behind a higher concentration of light elements. Thermal and compositional buoyancy at the inner core boundary (ICB) are more effective in driving convection than the secular cooling of the entire fluid core because they originate far from the rising fluid’s destination: the core–mantle boundary (CMB). Additional volumetric heating sources may exist in the fluid core due to radioactive decay, potassium-40 being the best candidate [Chabot and Drake, 1999; Buffett, 2000; Brodholt and Nimmo, 2002; Gasser and Wood, 2002; and Murthy et al., 2003]; the amount is uncertain and the focus of many current studies. Its presence
would have a significant effect on the age of the Earth’s solid inner core and the style of convection and magnetic field generation in the early Earth.

Hot, light fluid at the ICB, however, does not rise straight up to the CMB and likewise cold heavy fluid does not sink straight down. Coriolis forces (within the frame of reference of the rotating Earth) cause the low-viscosity fluid to flow in curved trajectories. The resulting twisting and shearing that this electrically conducting fluid does to the existing magnetic field converts kinetic energy into magnetic energy. Another way of saying this is that flow of metallic fluid through magnetic field produces an electromotive force that drives large electric currents (Ohm’s law). These electric currents induce new magnetic fields around them (Ampere’s law) that compensate for the continual decay of the magnetic field (Faraday’s law).

The fundamentals of dynamo theory, first proposed in the 1950’s, are the following: that differential rotation within the fluid core shears poloidal (north–south and radial) magnetic field into toroidal (east–west) magnetic field and that three-dimensional (3D) helical fluid flow twists toroidal field into poloidal field. At the same time, the more sheared and twisted the field the faster it decays away; that is, magnetic diffusion (reconnection) continually smoothes out the field. The field is self-sustaining if, on average, the generation of field is balanced by its decay. Discovering and understanding the details of how rotating convection in the Earth’s fluid outer core does this to maintain the observed intensity, structure and time dependencies requires 3D computer models of the geodynamo.

2. HISTORY OF DYNAMO MODELING

Many studies of magnetic field generation have been based on kinematic models. In these models the fluid flow induces magnetic field; but the feedback on the flow via Lorentz forces, which resist the movement of conducting fluid through the magnetic field, is neglected. Some compute or prescribe a large-scale 3D flow structure [e.g., Kumar and Roberts, 1975]. Other models compute or prescribe only the axisymmetric (two-dimensional) part of the flow and use it with a parameterized longitudinally-averaged effect of a hypothetical 3D helical flow to compute an axisymmetric magnetic field. Much has been learned from these studies about the types of fields that can be induced by various flow structures. However, the lack of the time-dependent magnetic feedback on the flow precludes a detailed prediction or explanation of the 3D flow and field structures in the Earth’s core, i.e., the geodynamo mechanism. In addition, based on rough estimates of the flow and field amplitudes in the core, the magnetic energy in the core is likely a thousand times greater than the kinetic energy of the flow that maintains it; therefore the neglect of Lorentz forces in kinematic dynamo models is not really appropriate for detailed investigations of the Earth’s dynamo mechanism.

Magnetohydrodynamic (MHD) dynamo simulations, on the other hand, are dynamically self-consistent, but also more challenging. They solve a coupled set of nonlinear differential equations that describe the 3D evolution of the thermodynamic variables, the fluid velocity and the magnetic field with all the major feedbacks. These geodynamo simulations are conducted to investigate the structure and time-dependence of convection and magnetic field generation in the Earth’s core. So little can be directly observed other than the surface magnetic field (today’s field in detail and the paleomagnetic field in much less detail) and what can be inferred from seismic measurements and variations in the length of the day and possibly in the gravitational field. Therefore, geodynamo models are used as much to predict what has not been observed as they are used to explain what has. When an MHD computer model generates a magnetic field with structure, intensity, and time dependence qualitatively similar to the Earth’s surface field, then it is plausible that the 3D flows and fields inside the model core are qualitatively similar to those in the Earth’s core. Analyzing this detailed simulated data provides a physical description and explanation of the model’s dynamo mechanism and, by assumption, of the geodynamo.

The first 3D global convective dynamo simulations were developed in the 1980’s to study the solar dynamo. Gilman and Miller [1981] pioneered this style of research by constructing the first 3D MHD dynamo model. However, they simplified the problem by specifying a constant background density, i.e., they used the Boussinesq approximation of the equations of motion. Glatzmaier [1984] developed a 3D MHD dynamo model using the anelastic approximation, which accounts for the stratification of density within the sun. Zhang and Busse [1988] used a 3D model to study the onset of dynamo action within the Boussinesq approximation. However, the first 3D MHD models of the geodynamo that successfully produced a dominantly dipolar field at the model’s surface were not published until 1995 [Kageyama et al., 1995; Jones et al., 1995; and Glatzmaier and Roberts, 1995a]. Since then several groups around the world have developed dynamo models [e.g., Kuang and Bloxham, 1997; Sarson et al., 1997; Kida et al., 1997; Busse et al., 1998; Christensen et al., 1998; Sakuraba and Kono, 1999; Katayama et al., 1999; and Hollerbach, 2000] and several others are currently being designed. Some features of the various simulated fields are robust, like the dominance of the dipolar part of the field outside the core. Other features, like the 3D structure and time dependence of the temperature, flow and field inside the core, depend somewhat on the chosen boundary conditions, parameter space and numerical resolution. Many review articles have been written that describe and compare these models [e.g., Hollerbach, 1996;
Although considerable progress has been made, none of these models has been able to run in a realistic parameter regime for the Earth's core. The problem is that, because of the low fluid viscosity, a broad spectrum of turbulence likely exists in the core, from global length scales down to scales on the order of meters; whereas current simulations have about 10 to 100 km resolution at best. In addition, because of the dominance of the Coriolis and Lorentz forces, this turbulence is heterogeneous and anisotropic. Capturing a good portion of this spectrum in a 3D geodynamo simulation would require more computing resources than are currently available. Therefore, greatly enhanced diffusion coefficients have been used in geodynamo simulations, which produces smooth large-scale (laminar) flows, not turbulence.

3. MODEL DESCRIPTION

We begin by briefly describing the Glatzmaier-Roberts geodynamo model, which employs the anelastic equations of motion. These equations allow for a realistic variation of density, temperature and pressure with depth, while filtering out acoustic waves. Effectively, the sound speed is assumed infinite so the pressure distribution at each numerical time step is in equilibrium throughout the fluid core. This approximation is valid when the fluid velocity is small relative to the local sound speed and the thermodynamic variations, or perturbations, are small relative to their background reference state values. These conditions are very well satisfied for the Earth's liquid core. The reason for using this approximation is that the numerical time step can be about a million times larger since fast-moving low-energy sound waves are not simulated. All other geodynamo models have employed the Boussinesq approximation, which assumes a constant background density. The small (20%) change in density across the Earth's fluid core would seem to justify this simpler approach; however, an important source of vorticity, which helps to generate magnetic field, is consequently neglected.

The anelastic MHD equations [Glatzmaier, 1984; and Braginsky and Roberts, 1995] that define our dynamo model describe the 3D, time-dependent perturbations of the flow, field and thermodynamic variables relative to a no-flow, non-magnetic, radially-dependent thermodynamic reference state fitted to the Preliminary Reference Earth Model (PREM). The set of equations ensures mass and magnetic flux conservation. An equation of state relates perturbations in the entropy, pressure and composition to density perturbations, which determine the perturbations in the gravitational field and the buoyancy forces. Newton's second law of motion determines how the local fluid velocity changes with time due to buoyancy, pressure gradient, viscous, Coriolis and Lorentz forces and to advection of momentum by the flow. The magnetohydrodynamic equations (i.e., Maxwell's equations and Ohm's law with the extremely good assumption that the fluid velocity is small relative to the speed of light) describe how the local magnetic field changes with time due to induction by the flow and diffusion due to finite conductivity. The second law of thermodynamics dictates how advection of entropy by the flow, thermal diffusion, and Joule and viscous heating determine the local time rate of change of entropy. A final advection-diffusion equation describes the local time rate of change of composition. These equations are solved each numerical time step to obtain the evolution of the 3D fluid flow and magnetic field and the perturbations in density, pressure, specific entropy, light constituent mass fraction and gravitational potential relative to the background reference state in a frame of reference rotating with the Earth's mean angular velocity.

The thermal boundary conditions at the ICB constrain the local flux of latent heat to be proportional to the local flux of light constituents and to the local cooling rate [Glatzmaier and Roberts, 1995]. The flux of light constituents through the CMB is set to zero. A non-zero heat flux is prescribed at the CMB; it controls the cooling rate of the core and therefore the production rate of buoyancy sources at the ICB and thus affects the intensity of the magnetic field. Normally, in this model the total heat flow out of the core is set to 7.2 TW, the generally assumed, although highly uncertain, value for the Earth. Of this, 5 TW is due to heat flow conducted down the adiabat. (Note, thermal convection occurs when the total temperature drop across the fluid outer core is greater than what it would be for an adiabatic temperature profile.)

The solid inner core and solid mantle rotate in reaction to magnetic torques, viscous torques if non-slip boundary conditions are applied, and gravitational torques if the gravitational forces between the mantle and the inner core are included [Buffett and Glatzmaier, 2000]. Viscous torques in the model represent the rate of momentum transfer between the fluid and the boundaries by small unresolved turbulence. The total angular momentum of the inner core, fluid core and mantle remains zero in the rotating frame, but the angular momentum of each of these is time dependent.

The magnetic field is generated in the model's fluid outer core and diffuses into its conducting solid inner core. Since the electrical conductivity of the Earth's mantle is very small relative to that of the core, everything above a thin layer at the base of the mantle is assumed to be a perfect insulator. The external magnetic field is therefore a potential field, albeit time dependent and determined by the dynamics inside the core.
The set of coupled nonlinear equations is solved at each numerical time step using a spectral method (spherical harmonic and Chebyshev polynomial expansions) \cite{Glatzmaier, 1984; and Glatzmaier and Roberts, 1996a}. The nonlinear terms are computed each time step by a spectral transform method, i.e., the simulated data are transformed from wave number space to grid space where the nonlinear products are computed, which are then transformed back to wave number space where the solution is advanced a time step. Simulations that span hundreds of thousands of years, involving tens of millions of numerical time steps, have been run at very low spatial resolution (e.g., 33 radial, 32 latitudinal, 64 longitudinal levels). Simulations have also been done at much higher resolution (e.g., 289 radial, 384 latitudinal, 384 longitudinal levels), which are more accurate because they resolve more of the energy spectrum; but they span less simulated time, on the order of tens of thousands of years.

The code is run on massively parallel computers with the data spread over the processors according to radial grid levels and spherical harmonic wave numbers. The spectral method requires each processor to send data to and receive data from every other processor every time step. This global communication among the parallel processors quickly grows as one increases the number of processors. In addition, although fast Fourier transforms are used for the spectral transform in longitude and radius, no efficient fast transform exists in latitude, and so this operation is very time consuming at high resolution.

The review articles mentioned above describe the variations on the equations and numerical methods employed in other geodynamo models \cite[e.g., Glatzmaier, 2002]{Glatzmaier, 2002}. Most dynamo models use spherical harmonic expansions, which have proven to be very accurate and efficient at relatively low spatial resolution. However, to reach much greater spatial resolution in the future, new methods are being developed that do not employ spherical harmonics and therefore avoid global inter-processor communication and the latitudinal spectral transform.

4. MODEL RESULTS

Since the mid 1990’s 3D computer simulations have advanced our understanding of the geodynamo. This is true even though they have been forced to use significantly enhanced diffusion coefficients that result in large-scale laminar convection. That is, the boundaries have a significant influence on the structure of the simulated flow because convective cells and plumes typically span the entire depth of the fluid outer core, unlike the turbulence that likely exists in the Earth’s core. The simulations have however shown that a strong, dominantly dipolar magnetic field, not unlike the Earth’s, can be maintained by convection driven by an Earth-like heat flux.

A typical snapshot of the simulated magnetic field from a geodynamo model is illustrated in Figure 1 with a set of field lines. In the fluid outer core, where the field is generated, field lines are twisted and sheared by the flow. The field that extends beyond the core is significantly weaker and dominantly dipolar at the model’s surface, not unlike the geomagnetic field. For most geodynamo simulations, the non-dipolar part of the surface field, at certain locations and times, propagates westward at about 0.2° per year, as has been observed in the geomagnetic field over the past couple hundred years \cite[e.g., Bloxham and Jackson, 1992]{Bloxham and Jackson, 1992}.

Several dynamo models have electrically conducting inner cores that on average drift eastward relative to the mantle \cite[e.g., Glatzmaier and Roberts, 1995a; Sakuraba and Kono, 1999; and Christensen et al. 2001]{Glatzmaier and Roberts, 1995a; Sakuraba and Kono, 1999; Christensen et al. 2001}, opposite to the propagation direction of the surface magnetic field. Inside the fluid core the simulated flow has a “thermal wind” component that, near the inner core, is predominantly eastward relative to the mantle. The magnetic field in these models that permeates both this flow and the inner core tries to drag the inner core in the direc-

![Figure 1](image)

Figure 1. A snapshot of the 3D magnetic field structure generated with the Glatzmaier-Roberts geodynamo model and illustrated with a set of magnetic lines of force. The axis of rotation is vertical and centered in the image. The field is a smooth, dipole-dominated, potential field outside the core. The color version of this is Plate 1 on the CD associated with this book; red represents outward directed magnetic field and blue represents inward directed.
tion of the flow. This magnetic torque is resisted by a gravitational torque between the inner core and mantle [Buffett and Glatzmaier, 2000].

Figure 2 shows the resulting time dependent angular rotation rate of the inner core relative to the mantle during a short interval of a simulation that prescribed a vanishing viscous torque on the inner core. Most of the time the inner core is rotating slightly faster than the mantle; but the rate is highly variable. The average super-rotation of the model’s inner core depends on the very poorly constrained viscosity assumed for the inner core’s deformable surface layer, which by definition is near the melting temperature. The amplitude of the super-rotation rate predicted by geodynamo models depends on the model’s specified parameters and assumptions; the original prediction was an average of about 2° longitude per year [Glatzmaier and Roberts, 1995a]. Since then the super-rotation rate of the Earth’s inner core today has been inferred from several seismic analyses [e.g., Song and Richards, 1996; Su et al., 1996; and Xu and Song, 2003], but is still quite uncertain [e.g., Creager, 1997; Souriau, 1998; and Laske and Masters, 1999]. Currently there is a spread in the inferred values, from the initial estimates of 1° to 3° eastward per year (relative to the Earth’s surface) to some that are zero to within an uncertainty of 0.2° per year. The average simulated rate in Figure 2 is closer to the lower end of this range; but again this depends on how deformable the model’s inner core surface is assumed to be.

On a much longer time scale, some dynamo simulations have produced spontaneous non-periodic magnetic dipole reversals [Glatzmaier and Roberts, 1995b; Sarson and Jones, 1999; Kageyama et al., 1999; Glatzmaier et al., 1999; and Kutzner and Christensen, 2002]. Periodic reversals, like the dynamo-wave reversals seen in early solar dynamo simulations [Gilman, 1983], have also occurred in recent dynamo simulations [Kida et al., 1997; Kida and Kittauuchi, 1998; Grote et al., 1999, 2000; and Simitev and Busse, 2002]. The highly variable times between reversals seen in the paleomagnetic record are measured in hundreds of thousands of years; whereas the time to complete a reversal is typically a few thousand years, less than a magnetic dipole decay time [e.g., Merrill and McFadden, 1999].

One of the simulated reversals is portrayed in Figure 3 with four snapshots spanning about 9000 years. The radial component of the field is shown at both the CMB and the surface of the model Earth. The reversal, as viewed in these surfaces, begins with reversed magnetic flux patches in both the northern and southern hemispheres. (It is interesting that a reversed flux patch is currently growing in the Earth’s southern hemisphere as the dipole moment of the geomagnetic field is slowly decreasing.) In addition, the longitudinally-averaged poloidal and toroidal parts of the field inside the core are illustrated at these times. Although when viewed at the model’s surface the reversal appears complete by the third snapshot, another three thousand years is required for the original field polarity to decay out of the inner core and the new polarity to diffuse in.

On an even longer time scale the frequency of reversals seen in the paleomagnetic record varies. The frequency of non-periodic reversals in geodynamo simulations has been found to depend on the magnitude of the convective driving relative to the effect of rotation [Kutzner and Christensen, 2002]. It also depends on the pattern of outward heat flux imposed over the CMB [Glatzmaier et al., 1999]. In the Earth, the CMB heat flux distribution might be determined by the distribution of cold subducted slabs at the base of the mantle. Since this distribution is determined by mantle convection, which is a million times slower than core convection, time-independent patterns of the CMB heat flux are prescribed via thermal boundary conditions. In addition, convection in the low viscosity outer core is so efficient that the maximum temperature variation in the core fluid at the CMB is no more than 10^3 K, compared with temperature variations in the high viscosity mantle of the order of hundreds of degrees K. Therefore the temperature drop, and so also the heat flux, between the core and mantle is presumably greatest where the mantle is relatively cold.

Two case studies are illustrated in Figure 4. One has higher heat flux imposed in the polar regions and in the equatorial region, with minimum heat coming out at mid-latitude. The other has a CMB heat flux patterned after today’s seismic tomography of the Earth’s lower mantle, with high heat flux along the high seismic velocity “rim around the Pacific” where cold slabs are thought to currently reside. The latitude of the south magnetic pole and the dipole moment are plotted for 300,000 years of these two simulations, for which the average numerical time step is about 15 days. The first case is seen to

![Inner core rotation rate relative to mantle](image)

**Figure 2.** The angular velocity of the solid inner core for a 20,000 year interval of a geodynamo simulation. Positive is eastward relative to the mantle. [From Buffett and Glatzmaier, 2000]
be extremely stable, with the magnetic pole never wondering far from the geographic pole. Since the natural convective heat flux within the fluid core preferentially transports more heat to the polar and equatorial regions, the imposed pattern of heat flux at the CMB for this case is very compatible with the internal fluid dynamics. Therefore magnetic instabilities have little chance of growing to significant amplitude. The second case is more Earth-like, with several reversal attempts and two successful reversals in the 300,000 years. Also, as seen in the paleomagnetic record, the intensity of the field typically decreases by at least an order of magnitude during a reversal.

Small changes in the local flow structure continually occur in this highly nonlinear chaotic system [e.g., Olson et al., 1999]. These can generate local magnetic anomalies that are reversed relative to what would be the direction of the global dipolar field structure. If the thermal and compositional perturbations continue to drive the fluid flow in a way that amplifies this reversed field polarity while destroying the original polarity, the entire global field structure would eventually reverse. However, more often the local reversed polarity is not able to survive and the original polarity fully recovers because it takes a couple thousand years for the original polarity to decay out of the solid inner core [Hollerbach and Jones, 1993; Glatzmaier and Roberts, 1995b]. This is a plausible explanation for “events” [Lund et al., 1998], which occur when the paleomagnetic field (as measured at the Earth’s surface) reverses and then reverses back, all within about ten thousand years [Gubbins, 1999; Glatzmaier et al., 1999].

These studies of reversals require long simulated times; and, other than setting the values of the diffusion coefficients and the pattern of heat flux out of the CMB, they are not adjusted to get preferred results. Also, the simulated reversals are not externally triggered; they occur naturally and spontaneously in this highly nonlinear system. However, these long simulated times require tens of millions of numerical

Figure 3. 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 core–mantle boundary and at what would be the surface of the Earth, displayed at roughly 3000-year intervals spanning a dipole reversal from a geodynamo simulation. In the plots of the average field, the small circle represents the inner core boundary and the large circle is the core–mantle boundary. The poloidal field is shown as magnetic field lines on the left-hand sides of these plots (light shade or blue is clockwise and dark or red is counter-clockwise). The toroidal field direction and intensity are represented as contours (not magnetic field lines) on the right-hand sides (dark or red is eastward and light or blue is westward). Aitoff-Hammer projections of the entire core–mantle boundary and surface are used to display the radial component of the field (with the two different surfaces displayed as the same size). Light shades or red represent outward directed field and dark or blue represent inward field; the surface field, which is typically an order of magnitude weaker, was multiplied by 10 to enhance the color contrast. [From Glatzmaier et al., 1999] The color version of this image (Plate 2) is on the CD associated with this book.
time steps, which can only be afforded at relatively low spatial resolution. Alternatively, studies of the structure and time dependence of the flow and field during relatively quiet times between reversals requires less simulated time and therefore can be done at much greater spatial resolution.

For example, in the 1970s, there was a debate concerning the presence of radiogenic elements in the Earth's core. Due to a lack of convincing evidence to support their presence, this debate ended with a general consensus that there are no significant amounts of radioactive elements present [e.g., Stacey, 1992]. Recently, however, new evidence in the form of experimental studies [e.g., Chabot and Drake, 1999; Gessman and Wood, 2002; and Murthy et al., 2003] and thermal evolution models [e.g., Buffett, 2000; and Brodholt and Nimmo, 2002] have shown that potassium-40 may have partitioned into the Earth's iron core during its formation; however, the predicted concentration today is still debated, ranging between 1 ppm and 300 ppm.

In an attempt to shed some light on this question, we have run a modified version of the Glatzmaier-Roberts geodynamo model to check if there are any significant effects on the simulated magnetic field due to radiogenic heating from 250 ppm of potassium-40 in the core. This corresponds to prescribing 25% of the core–mantle boundary heat flow to be coming from a volumetric heating source proportional to the radially-dependent density. Another case, the control case, was run without radiogenic heating. The other three quarters of the CMB heat flow for the internally heated case (and the total for the control case) comes from the latent heat source at the ICB, Joule and viscous heating and secular cooling. Otherwise the two cases are the same. In this model, the viscous, thermal, compositional and magnetic diffusivities are set to be constant, equal and independent of the length scale of the spherical harmonic mode. (Note, this prescription differs from previous Glatzmaier-Roberts simulations and the calculations have been performed at much greater spatial resolution.)

Snapshots of the vertical flow in the equatorial plane are shown in Figure 5. Notice that even for these more highly resolved and less diffusive simulations, the convective plumes typically span the entire depth of the outer core. That is, these simulations are still not turbulent. The results show a subtle difference in the vertical flow structure (Figure 5): the internally heated case has weaker upwellings at the ICB. This occurs because the thermal gradient at the ICB, which drives convection there, is less steep relative to the control case since the diffusive heat flux (and compositional flux) there is less. Note, that this also implies that the inner core for the internally heated case would be much older than the same size inner core of the control case.

The magnetic field structures for the two cases look surprisingly similar; a snapshot of the field for the internally heated case is displayed in Figure 1. The internally heated case does, however, maintain an average magnetic field 10–20% more intense than that of the control case, and an average kinetic energy 10–20% less (Figure 5) than that of the control case. There are also slight differences in the magnetic energy spectra between the two cases. The internally heated case, unlike the control case, usually has a greater quadrupole component than octupole at the surface, as is the case for the present day Earth. The control case however also displays this characteristic occasionally. Unfortunately, we have seen no significant difference in the surface fields of the two cases to argue for or against potassium-40 in the Earth's core. This is in agreement with a similar dynamo study [Kutzner and Christensen, 2002]. Higher spatial resolution and lower diffusivities may be needed to produce a noticeable difference in the dynamo.

Many other studies have been conducted via dynamo simulations to, for example, assess the effects of the size and conductivity of the solid inner core [e.g., Sakuraba and Kono, 1999; Bloxham, 2000; Morrison and Fearn, 2000; Roberts and Glatzmaier, 2001; and Wicht, 2002], of a stably stratified layer at the top of the core [Glatzmaier and Roberts, 1997], of heterogeneous thermal boundary conditions [e.g., Sarson et al., 1997; Glatzmaier et al., 1999; Bloxham, 2000;
Olson and Christensen, 2002; and Christensen and Olson, 2003], of different velocity boundary conditions [e.g., Kuang and Bloxham, 1999; Kuang, 1999; and Christensen et al., 1999] and of computing with different parameters [e.g., Gilman, 1983; Olson et al., 1999; Christensen et al., 1999, Grote et al., 2000; Simitev and Busse, 2002]. These models differ in several respects. For example, models employ either the Boussinesq or the anelastic approximation, compositional buoyancy is usually neglected as are perturbations in the gravitational field, different boundary conditions and spatial resolutions are chosen, the inner core may be treated as an insulator instead of a conductor or may not be free to rotate. As a result, the simulated flow and field structures inside the core differ. For example, the strength of the shear flow on the “tangent cylinder” (the imaginary cylinder tangent to the inner core equator), which depends on the relative dominance of the Coriolis forces, varies among the simulations. Likewise, the vigor of the convection and the resulting magnetic field generation tends to be greater outside this tangent cylinder for some models and inside for others. But all the solutions have a westward zonal flow in the upper part of the fluid core and a dominantly dipolar magnetic field outside the core.

However, when assuming Earth values for the radius and rotation rate of the core, all models of the geodynamo have been forced (due to computational requirements) to use a greatly enhanced viscous diffusivity to account for mixing by the small-scale (unresolved) turbulence. However, the values used have been at least three or four orders of magnitude larger than estimates of what it should be for the current spatial resolutions that are employed, which is roughly equal to the Earth’s actual magnetic diffusivity. One must also decide how to prescribe the thermal, compositional and magnetic diffusivities [e.g., Dormy et al., 2000; Kono and Roberts, 2002; Simitev and Busse, 2002; and Glatzmaier, 2002]. They all have units of length squared per time and serve as coefficients that relate the degree of structure in the temperature, composition and magnetic field, respectively, to the rates these quantities diffuse. One of two extremes has typically been chosen. These three diffusivities could be set equal to the Earth’s actual magnetic diffusivity (2 m²/s), making the viscous diffusivity much greater than these; this was the choice for the previous Glatzmaier-Roberts simulations. Alternatively, they could be set equal to the enhanced viscous diffusivity, making all (turbulent) diffusivities too large, but at least equal; this was the choice of most of the other models, including the internally heated version of the Glatzmaier-Roberts model presented above. Neither choice is satisfactory.

5. TWO-DIMENSIONAL MODELS

The fundamental question about geodynamo models is how well do they simulate the actual dynamo mechanism of the Earth’s core. Many of us have argued, or at least suggested, that the large (global) scales of the temperature, flow and field seen in these simulations should be fairly realistic because the diffusivities may be asymptotically small enough. That
is, although molecular viscosity of the fluid in the outer core may be as much as $10^{10}$ times smaller than what the (turbulent) viscosity is set to in current geodynamo simulations, viscous forces in most simulations (away from the boundaries) still tend to be about $10^4$ times smaller than the Coriolis and Lorentz forces. No one believes geodynamo simulations need to use the actual molecular viscosity to be able to get realistic large-scale flows because the transport of momentum by the small unresolved turbulence is much more efficient than actual viscous forces. All models of the geodynamo have in fact crudely modeled this nonlinear mixing process as a simple linear diffusion process with a very enhanced viscous diffusivity. The question is how much of the turbulence spectrum needs to be numerically resolved to depict adequately the large scale flow, and therefore field, in the Earth’s core, i.e., the geodynamo mechanism. Only when numerical methods and computing resources improve to the point where we can further reduce the turbulent diffusivities by several orders of magnitude and produce at least moderately turbulent 3D dynamo simulations will we be able to answer this fundamental question.

To gain some insight to what the answer might be, we examine two simulations of two-dimensional (2D) magnetoconvection. Magnetoconvection refers to an MHD simulation with an imposed background magnetic field; convection amplifies and distorts the field and Lorentz forces affect the flow, but, unlike in a convective dynamo, the field cannot decay away. Although the 2D constraint precludes realistic global flows and a dynamo mechanism, it does allow us to run at much greater spatial resolution and therefore with much smaller viscous, thermal and magnetic diffusivities. That is, we are able to run in a more realistic parameter regime, one that we would like to be able to reach someday in 3D.

We specify a box of fluid in the horizontal ($x$) and vertical ($z$) directions, with no flow, field or variations in the $y$ direction. There is a constant gravitational acceleration in the $-z$ direction; and the frame of reference is rotating about an axis in the $y$ direction. This box can be considered a small part of the equatorial plane, far from the axis of rotation. The top and bottom boundaries are impermeable and stress free and there is a specified drop in specific entropy across the depth of the box. There is an externally applied, uniform, background magnetic field in the $z$ direction. The side boundaries are periodic.

We solve a system of magnetohydrodynamic equations in 2D for the thermodynamic perturbations, the fluid flow and the induced magnetic field using a numerical method that has been employed in many previous 2D Boussinesq studies [e.g., Weiss, 1981], but with modifications that account for the density stratification of our anelastic model [Rogers et al., 2003; and Glatzmaier, 2004]. The numerical time step is limited by a Courant condition based on the fluid and Alfvén velocities and the grid resolution. That is, to maintain numerical stability, the time step cannot exceed the shortest time it takes fluid to flow, or magnetic waves to propagate, between any two adjacent grid points. We use 2001 (non-uniform) grid levels in the vertical direction and 4001 (uniform) grid levels in the horizontal direction.

We begin by describing a 2D simulation that is time dependent but laminar. The density stratification across the convection layer is set to that of the Earth’s outer fluid core, i.e., the density at the bottom boundary is about 20% greater than that at the top. The viscous, thermal and magnetic diffusivities are constant and equal but much larger than Earth values. Buoyancy, rotational, magnetic and viscous effects on the flow are similar in magnitude to those in current 3D geodynamo simulations.

The resulting convective velocities are a thousand times greater than the diffusive velocity and ten times smaller than the rotational velocity. The induced magnetic field has an average maximum intensity about an order of magnitude

![Laminar convection](image1)

![Turbulent convection](image2)

**Figure 6.** Snapshots of the entropy for two different 2D anelastic, rotating, magnetoconvection calculations. Gravity is downward. Hot (red) plumes rise from the bottom boundary and cold (blue) sink from the top. The turbulent case has viscous, thermal and magnetic diffusivities one thousand times smaller than those of the laminar case. The color version of this image is Plate 4 on the CD associated with this book.
greater than the applied background field intensity and the magnetic energy is comparable to the kinetic energy. Figure 6 is a snapshot of the perturbation of entropy (approximately the temperature perturbation). The flow is said to be “laminar” because thermal plumes extend from one boundary to the other, similar to large convection cells. The evolution resembles the flow seen in a lava lamp. This time-dependent large-scale laminar convection is typical of the style of convection simulated with current 3D geodynamo models.

Now we examine what happens when we reduce all three diffusivities, each by a factor of a thousand. Convective velocities are now almost a million times greater than the diffusive velocity and the resulting Coriolis forces are typically 10^9 times greater than the (turbulent) viscous forces, closer to the conditions in the Earth’s fluid core. A snapshot of this turbulent case is also shown in Figure 6. As expected, there is much more energy in the small spatial scales compared with the laminar case. After detaching from the boundaries, turbulent plumes interact with each other, not the boundaries.

However, somewhat unexpected, deep turbulent boundary layers exist. This occurs because kinetic energy of buoyant plumes rising from the bottom boundary is converted into rotational kinetic energy. That is, as plumes rise from the bottom boundary the Coriolis forces resulting from the expansion (due to the slight density stratification) generate vortices with counter-clockwise rotation. Clockwise rotating vortices are generated in plumes sinking from the top boundary. This effect is relatively insignificant in the laminar case and in the current 3D anelastic geodynamo simulations because viscosity in these calculations is too large. It is completely absent in Boussinesq simulations, which do not account for density stratification.

A movie of this turbulent case shows high-frequency high-amplitude Alfvén waves manifested as local horizontal oscillations of the entropy and magnetic field structures. The kinetic energy of this wave motion exceeds the kinetic energy of the convection. This effect is not seen in current 3D geodynamo simulations because the large viscosity used in those simulations damps these waves.

6. NEXT GENERATION OF MODELS

Our understanding of how the geomagnetic field is generated by convection in the Earth’s outer fluid core has improved during the past decade with the development of magnetohydrodynamic dynamo models. The relatively Earth-like magnetic field structure and time dependence, including dipole reversals, produced by these models is certainly encouraging. However, no dynamo model has yet simulated strongly turbulent convection, which likely exists in the Earth’s core. We have assumed that the enhanced diffusivities in our laminar simulations are adequately, albeit crudely, accounting for the mixing by the unresolved turbulence. Two-dimensional magnetoconvection simulations, however, demonstrate that strongly turbulent rotating magnetoconvection is significantly different than the corresponding laminar simulations obtained with much larger diffusivities.

These findings suggest that current 3D laminar dynamo simulations may be missing critical dynamical phenomena. It is important, therefore, to strive for much greater spatial resolution in 3D models in order to significantly reduce the enhanced diffusion coefficients and actually simulate turbulence. This will require faster parallel computers and improved numerical methods, which hopefully will happen within the next decade or two. In addition, sub-grid scale models need to be added to geodynamo models to better represent the heterogeneous anisotropic transport of heat, composition, momentum and possibly also magnetic field by the part of the turbulence spectrum that remains unresolved.

How geophysically realistic is the laminar dynamo mechanism in the current geodynamo models? Whatever the answer, these findings suggest that there are still several exciting and important phenomena about the geodynamo waiting to be discovered with the next generation of models.

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REFERENCES


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