Lectures in Paleomagnetism, 2005
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Chapter 14

The ancient geomagnetic field

Suggested Reading

For background:
Chapters 1 and 7: Butler (1992)
Lecture 2: this lecture series
To learn more:
Chapters 1, 2.4, 4, 6.4: Merrill et al. (1996)

14.1 Motivation for study of the behavior of the Earth's magnetic field

The magnetic field is one component of the highly complex Earth system. It interacts with the atmosphere, the biosphere, the deep mantle and even the inner core. It also has the useful property of pointing roughly North (or South). Records of the Earth's magnetic field play a role in many aspects of Earth Science; hence some knowledge of how it behaves is important to all Earth scientists. The following is a partial list of reasons to study the geomagnetic field

• Atmospheric interaction

Radioactive forms of carbon, beryllium and chlorine are produced in the atmosphere by cosmic ray bombardment. The decay of these isotopes is used for dating purposes in a wide variety of disciplines. There are large variations in ages predicted from tree ring, varve or ice layer counting or U/Th dating and those estimated by radiocarbon dating (see Figure 14.1). Some of these variations could be caused by changes in the carbon balance between the atmosphere and the deep ocean (which is a reservoir of old carbon) and some could be caused by changes in magnetic field strength. Because the magnetic field shields the atmosphere to a large extent from cosmic rays, changes in the intensity of the magnetic field result in changes in production, hence are a key parameter in deriving accurate age information. To date, there is rather poor agreement between the variations in radiocarbon production predicted using changes in paleointensity of the geomagnetic field (see, e.g., Figure 14.1.) Either the field variations are not known, the relationship between those variations and radiocarbon production is not known or the actual variations in production are not known because of unconstrained reservoir effects (or any combination of these factors).

• Biospheric interaction

Some living forms make magnetic crystals (Lecture 6). In the case of magnetotactic bacteria, these tiny magnets are used for physical orientation. In some cases, animals may use magnetic field

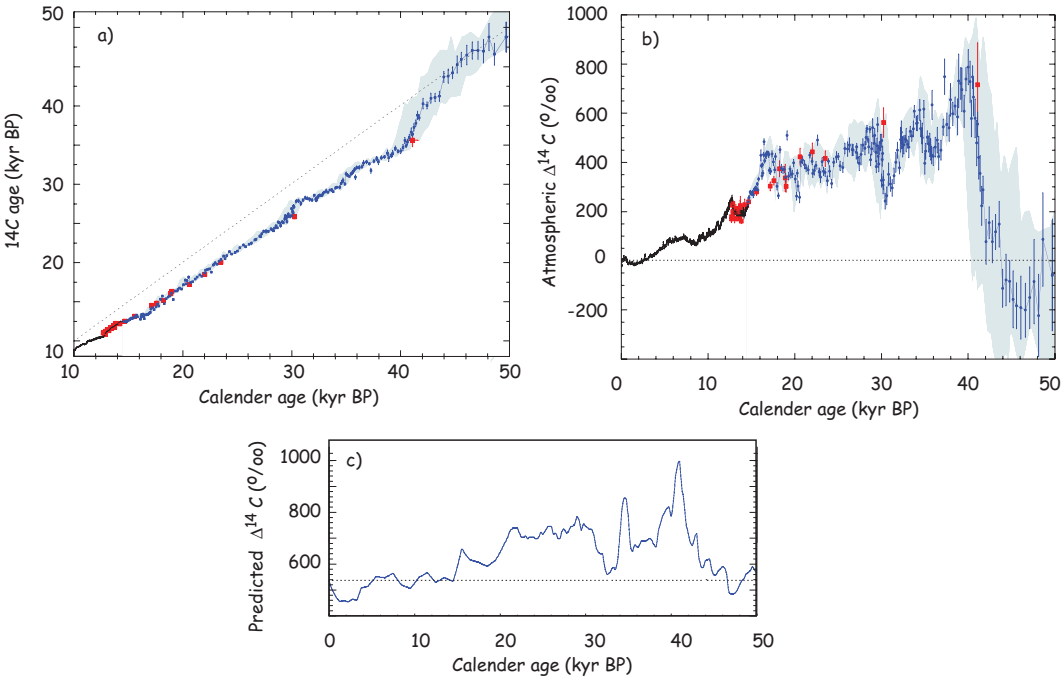


Figure 14.1: a) Radiocarbon calibration data from from Cariaco ODP Leg 165, Holes 1002D and 1002E (blue circles), plotted versus calendar age assigned by correlation of detailed paleoclimate records to the Greenland Ice Core GISP2. The thin black line is high-resolution radiocarbon calibration data from tree rings joined at 12 cal. ka B.P. to the varve counting chronology. Red squares are paired ^{14}C-U/Th dates from corals. Light gray shading represents the uncertainties in the Cariaco calibraion. The radiocarbon dates are too young, falling well below the dashed line of 1:1 correlation. b) Compilation of data interpreted as production rate changes in radiocarbon (Δ^{14}C) versus calender age. (symbols same as in a). c) Predicted variation of Δ^{14}C from the geomagnetic field intensity variations from sediments of the north Atlantic (Laj et al., 2002) using the model of Masarik and Beer (1999). [Figure modified from Hughen et al., 2004.]

lines for navigation.

- Deep mantle interaction

Studies of seismic waves have demonstrated large variations in seismic velocity near the core mantle boundary. There appears to be an annulus of faster velocities surrounding the Pacific ocean which may reflect the influence of cold subducted slabs. The geomagnetic field is generated by convection in the outer core. This convection could be a strong function of the thermal gradient at the top of the core. Temperature variations therefore could conceivably have an effect on the geomagnetic field (e.g., Glatzmaier et al., 1999). Is there any evidence for this? Are there any changes in the magnetic field as a function of long term changes in the CMB?

- Inner core interaction

Numerical simulations of the magnetic field predicted that the process of generation of the magnetic field interacted with the inner core in such a way as to make it spin faster than the rest of the Earth (Glatzmaier and Roberts, 1996). The effect has been sought in seismic data (e.g., Song and Richards, 1996), although its existence is still a matter of debate.

• Tectonic and Geological applications

Paleomagnetic data often are a critical component of stratigraphic and tectonic investigations because they provide temporal and paleogeographic constraints unavailable by any other method. Therefore, it is useful to know what sorts of data can be expected from records of the geomagnetic field, as oppose to geological modification through overprinting or post-formation rotation. It is also useful to know how long one must average the observations to achieve a reasonable estimate of the time averaged field (TAF) and whether or not it can be approximated by a GAD model.

•The sky is falling!

Hulot et al. (2002) were among the first to point out the fact that the Earth's magnetic field has dropped in intensity over the last two decades. This observation, combined with the fact that the reverse flux patches on the core mantle boundary appear to be growing lead to speculation that the geomagnetic field might be starting to reverse its polarity. What is the likelihood that this will happen? What does the field do when it is about to reverse? What is the average intensity of the field and how frequently does it do what it is doing now without reversing?

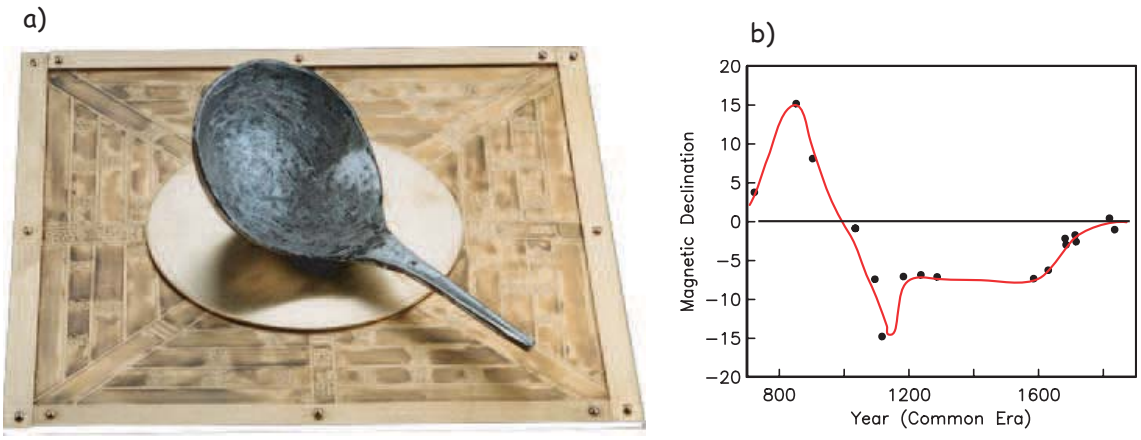


Figure 14.2: a) A reconstruction of the south pointing spoon (*shao*) used by the Chinese in the second century BCE. b) Measurements of magnetic declination made in China from 720 CE to 1829. [Compiled by Smith and Needham, 1967.]

In order to answer some of the questions just raised, we need measurements of the geomagnetic field. The geomagnetic field changes on frequencies of 10s of microseconds (radio waves) to millions and perhaps billions of years. Direct observations contribute to our knowledge of field behavior up to the order of 100s of years, but on longer times scales we need to use paleomagnetic and archeomagnetic techniques. We will first review what is known from historical measurements of the geomagnetic field. Then we will turn to what we can glean from "accidental records" made by archeological and geological materials.

14.2 Historical measurements of the Earth's Magnetic field

The magnetic properties of lodestone were already well known by at least several centuries BCE. Legend has it that a boy in ancient Greece discovered the magnetic properties of lodestone that he found near the city of Magnesia. The earliest compass dates from the second century BCE. Lodestone spoons (see Figure 14.2a) were placed on bronze plates, often decorated with images of

the Big Dipper and other heavenly images. These "south pointers" were apparently used primarily for prognostication, geomancy and Feng Shui. It was not until sometime in the late 14th Century that compasses were used for sea-going navigation in China.

That the magnetic field changes was apparently first discovered in China around 720 CE when the astronomer Yi-Xing measured magnetic declination (see Figure 14.2b). The compass did not arrive in Europe until some time in the 12^{th} century, but the idea of declination did not accompany it. The deviation of magnetic north from true north was not rediscovered until the early 1400s. Europeans began to make systematic measurements of declination in the early 1500s. Magnetic inclination was discovered in the mid-1500s in Europe.

Gilbert (1600) noted variations in field strength with latitude based on the sluggishness or rapidity with which a compass settled on the magnetic direction. Magnetic intensity was first measured quantitatively in the late 1700s by French scientist Robert de Paul, although all records were lost in a ship wreck. The expedition sent to search for the lost ship made several measurements, using the period of oscillation T of a vertical dip needle with magnetic moment m and moment of inertia I. These are related to B by:

$$T = 2\pi\sqrt{\frac{I}{mB}}.$$

These measurements supported Gilbert's hunch that the intensity of the field increases away from the equator.

The internal origin of the magnetic field was discovered by Gilbert in 1546 who made a systematic study of the magnets and the the Earth's magnetic field, published in 1600. While aware of deviations of magnetic declination from true north, Gilbert thought that the field was unchanging in time. It wasn't until Gellibrand, in 1634, compared declination measurements made in London over a period of some 50 years and concluded that the geomagnetic field changes. Thus Europeans discovered secular variation of the magnetic field in 1634, over a millenium after the Chinese.

Scientific exploration at sea began with the expeditions of the Pink Paramore under Caption Edmond Halley (1698-1701). Halley published the first geomagnetic chart in 1702 (see Figure 14.3). He noticed that some geomagnetic features appeared to be moving to the west, a phenomenon known as *westward drift*. compare for example the "line of no variation" in Figure 14.3 with the line of zero declination in Figure 2.5 in Lecture 2. It has moved significantly to the west in the equatorial and southern Atlantic realms.

Gauss provided the mathematical framework we use today to deal with geomagnetic data when he invented the spherical harmonic expression for the geomagnetic potential field (see Lecture 2). The first such analysis (done in 1838) was based on 84 data points evaluated on an evenly spaced grid from isomagnetic charts of the magnetic field elements available at the time.

Fastforwarding to our own millenium, we find researchers still poring over these centuries old measurements. Bloxham, Gubbins, Jackson and colleagues are mining the maritime records of early European sea-faring merchants. These ship's logs contain a huge treasure trove of measurements of declination and sometimes inclination since the 16th century (see e.g., Jackson et al., 2000). The intensity of the radial component of the magnetic field inferred for the core mantle boundary at two time intervals is shown in Figure 14.4. Compare Figure 14.4b with Figure 2.5a in Lecture 2 which is the intensity of the magnetic field observed at the surface. There are more so-called "flux patches" (the spots of higher intensity) in Figure 14.4b because the field was evaluated closer to the source (the core), but the general pattern is similar. The field for 1600, however, was somewhat



Figure 14.3: Chart of magnetic declination from Halley, (To make observations on the magneticall needle etc.: the Voyages of the Pink Paramore under captain Edmond halley, FRS: 1698-1701) published in 1702.

different. The number and positions of the flux patches has changed substantially since then. Some flux patches, in particular the prominant patch that is now over Africa, have drifted westward from the Indian ocean, a phenomenon largely responsible for "westward drift".

As already mentioned, observatory measurements of the intensity of the magnetic field are only available since the mid-19th century. These show that the large changes in declination and inclination were also accompanied by even more dramatic changes in field strength. We plot the intensity of the field evaluated from the **gufm1** model of Jackson et al. (2000) for San Diego, CA, in Figure 14.4c. If the field continues on its recent trajectory, it will reach zero by the year 2500.

14.3 The archeo- and paleomagnetic record of geomagnetic field behavior

Historical observations quickly run out as we go back in time. Prior to 720 CE there are no surviving human measurements. Yet the average field based on the historical measurements (e.g., Jackson

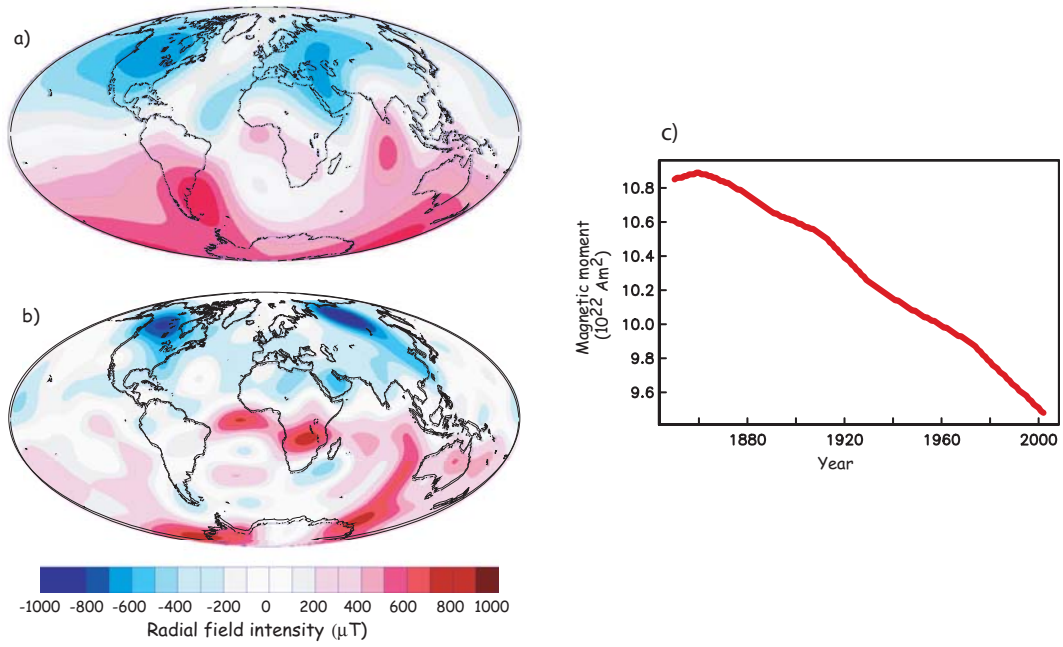


Figure 14.4: Maps of the intensity of the radial magnetic field at the core mantle boundary. b) Deduced from the secular variation measurements compiled by Jackson et al., (2000) for 1600. b) Map from the IGRF of 1990. c) Magnetic intensity in San Diego, CA evaluated from the **gufm1** model.

et al., 2000) is anything but GAD. To see how observations of the magnetic field such as westward drift, quasi-stationary flux lobes and the degree of "GADness" change through time, we must turn to rock and archeological materials to give us a picture of the ancient geomagnetic field.

14.3.1 Pioneers in paleomagnetism

Strongly magnetized rocks (as opposed to the mineral lodestone) had been noticed during the 1700s because of their effect on compass needles, but the fact that certain rocks were magnetized in the direction of the Earth's field was discovered by Delesse in 1849 and Melloni in 1853. Folgerhaiter extended the study of fossil magnetizations to the magnetic properties of baked archeological materials in 1899. Naturally baked material (heated by lava flows) was studies by David (1904) and Brunhes (1906). In the course of their investigations, they discovered materials that were magnetized in a direction opposite to the Earth's field. This first application of the baked contact test led to speculation that the Earth's field had reversed its polarity in the past.

Mercanton (1926) argued that the field had reversed based on the fact that reversely magnetized rocks were found all over the world. Matuyama (1929) further supported the argument by demonstrating that all the reversely magnetized rocks in Japan were older than the overlying normally magnetized rocks. It was not until the combined use of paleomagnetism and K-Ar dating allowed researchers in the U.S. and Australia (e.g., Cox et al., 1963) to demonstrate the global synchrony of polarity intervals that the scientific community embraced the notion of polarity reversals.

Sedimentary materials were first used for the investigation of secular variation by Johnson et al. (1948) who measured samples from varved lakes in New England. Mackereth developed a

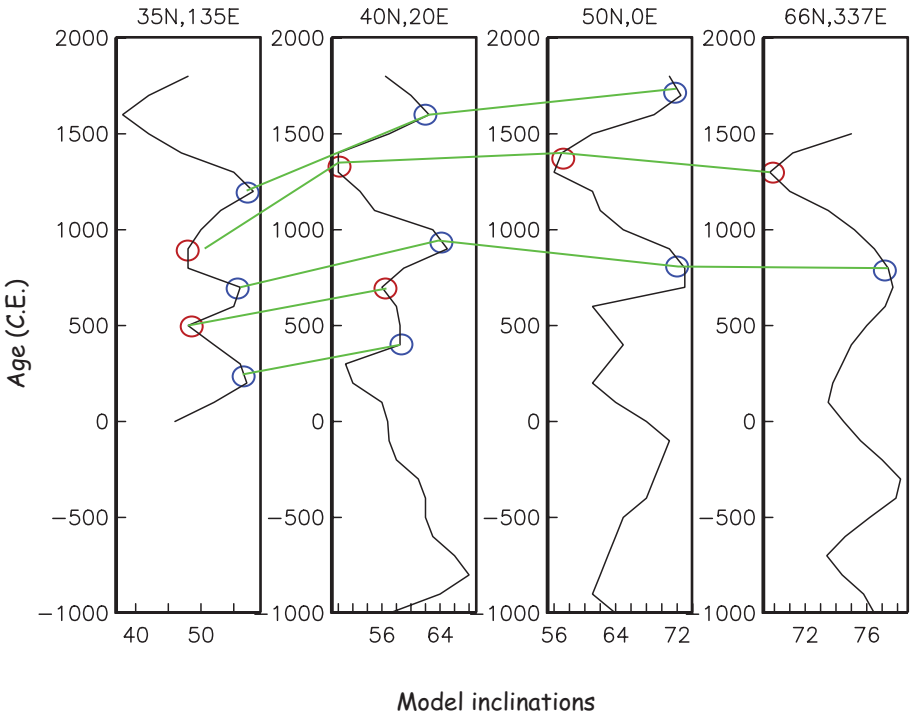


Figure 14.5: Inclinations evaluated at 100 year intervals from the PSVMOD1.0 of Constable et al. (2000) for selected records. These are plotted from West to East. Maxima and minima are noted. Westward drift would imply that these features would "rise" to the right.

pneumatic for coring device for use in lakes in 1958, opening the way for studies of the detailed time variations of the magnetic field.

14.3.2 The last few millenia

Constable et al. (2000) assembled a data set of 24 time series of directional data from archeomagnetic and lake sediment sources evaluated at 100 year intervals (PSVMOD1.0). We plot examples of several of the inclination records from West to East in Figure 14.5. The most recent attempt at spherical harmonic modelling of archeomagnetic, marine and lake sedimentary data is that of Korte and Constable (2005) who have pressed back 7000 year and have included intensity data in their modelling efforts.

14.3.3 Westward drift

We mentioned that early workers measuring the secular variation of declination noticed that certain features appeared to move west with time. A careful look at the data shows that this tendency as a subtle, probably only locally observed effect. Yukutake (1967) collected together the data available at the time and marked the occurrences of maxima and minima in both declination and inclination. Some of these are marked on Figure 14.5 as examples. Yukutake then plotted these maxima and minima as a function of age and longitude of the observation site. The data appeared to suggest that the features moved westward at a rate of about a half a degree per year. This would

mean that the maxima and minima on Figure 14.5 would rise to the right as they sort of do, but the data are rather unconvincing.

14.3.4 The more distant past

For more distant times in the past, we look at records from one place over time. In Figure 14.6, we see an example of one such record, obtained from dry lake sediments exposed along the shores of Mono Lake. The geomagnetic field oscillated around the GAD direction with an amplitude of some 30° over an interval of some 9 meters (the age is debated, but recent estimates place the record at approximately 38-41 ka; Kent et al., 2002) On rare occasions, the field departs drastically from what can be considered normal of secular variation and executes what is known as a *geomagnetic excursion*. The definition of a geomagnetic excursion is usually magnetic records in which the VGPs are more than 45° away from the average pole for that time and place. The convention is to name the excursion after the place where it was first observed, so this would be the "Mono Lake excursion", if it is distict from the "Laschamp excursion" discovered in volcanics near Laschamp, France.

Excursions are thought to be accompanied by a decreases in paleointensity (DIPs) (a paleointensity low). For this reason, some recent studies have identified "excursions" based on the occurence of paleointensity lows (see Figure 14.7), but this is strictly speaking incorrect.

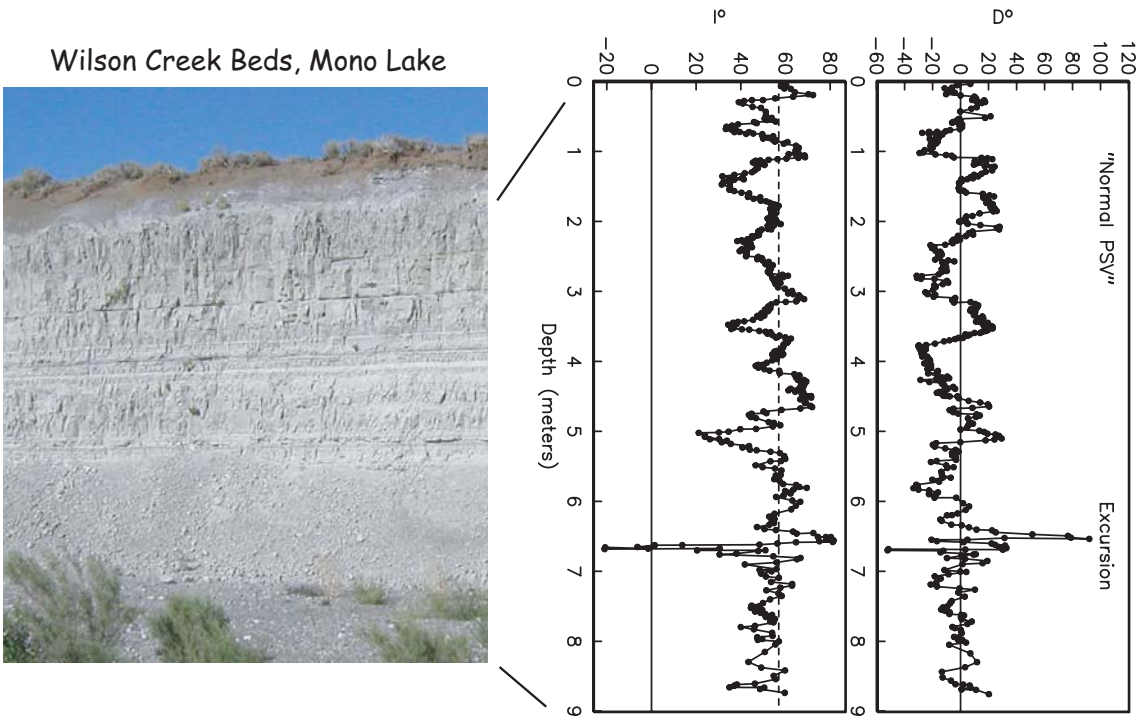


Figure 14.6: Paleosecular variation of the magnetic field (D and I) observed in the Wilson Creek section north of Mono Lake. The inclination expected from an geocentric axial dipole is shown as a dashed line. The declination is expected to be zero. The so-called "Mono Lake" excursion is marked. The data are from Lund et al. (1988) and represent some 23 kyr of time.

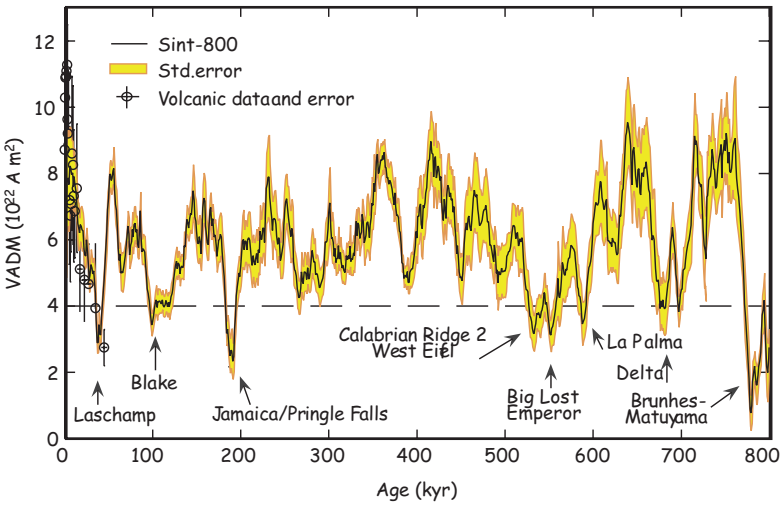


Figure 14.7: Stack of relative paleointensity records from deep sea sediments. [Figure modified from Guyodo and Valet, 1999.]

When viewed over sufficient time, the geomagnetic field reverses its polarity, by which we mean that the sign of the axial dipole term (g_1^0) changes. An example of a so-called *polarity reversal* is shown in Figure 14.8a (Clement and Kent, 1984). The intensity of the magnetic field appears to drop to approximately 10% of its average value and the directions migrate from one pole to the other over a period of several thousand years. When the polarity is the same as the present polarity it is said to be *normal*. When it is in the opposite state, it is said to be *reverse*. The duration of the reversal process also appears to be a function of latitude (Clement, 2004).

The details of what happens during a polarity reversal are still rather unclear because they occur so quickly, geologically speaking. Some high resolution sedimentary records are like that shown in Figure 14.8 whereby there is an orderly progression from one polarity to the other. However, a polarity transition captured by rapidly erupted lava flows records a more complex picture (see Figure 14.8b). There are a few conclusions we can draw however: 1) they occur quickly and 2) they are always associated with low geomagnetic intensities (see Figure 14.9a).

A more controversial observation about directions in extrema was first pointed out by Clement (1991): when mapped to VGP positions, they plot in preferred paths (see Figure 14.9b). These paths are seen in most data sets, but can be made to disappear when certain criteria are applied (e.g., Prévot and Camps, 1993). The intriguing thing about the paths is that they appear to coincide with the shear velocity anomalies (see Figure 14.9c). Whether or not the paths exist has been debated ever since. However, recent work by Love (1999) seems to support the preferred path hypothesis.

On average, the field spends about half its time in each polarity state, and only a tiny fraction (1-2%) of the time in an intermediate state. Rocks of both polarities have been documented from early in the Earth's history (at least since the late Archean, see Strik et al. 2003), although the frequency of reversal has changed considerably through time (see Opdyke and Channell, 1996 and Merrill et al., 1996).

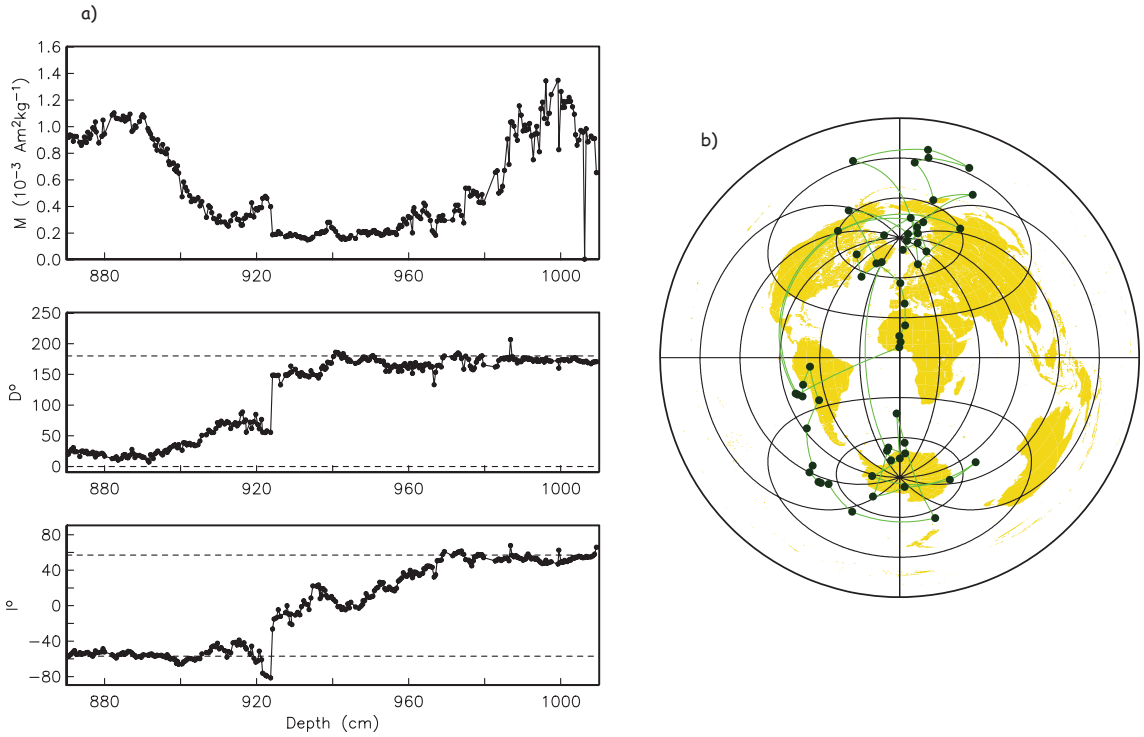


Figure 14.8: a) The lower Jaramillo geomagnetic polarity reversal as recorded in deep sea sediments from core RC14-14. Inclinations and declinations expected from a normal and reverse GAD field are shown as dashed lines. [Data from Clement and Kent, 1984]. b) Record of polarity transition recorded at Steens Mountain. [Data from Camps et al., 1999.]

14.4 Geomagnetic polarity time scale - a first look

A list of dates of past geomagnetic polarity reversals is known as a *geomagnetic polarity time scale* (GPTS). How the time scale is calibrated is discussed in the next lecture. For now we will just take it as a given. In Figure 14.10 we show the polarity history from the marine magnetic anomaly template. The details of the history of reversals for times older than the oldest sea floor magnetic anomaly record (about 160 Ma) are sketchy, but will eventually be documented using sedimentary records of the magnetic field (see e.g., Kent and Olsen, 1999).

Examination of the reversal history shown in Figure 14.10 suggests that reversals occur at apparently random intervals without a predictable pattern. Furthermore, the frequency of reversals appears to change (see for example, Constable, 2003). Above the polarity history in Figure 14.10, we plot the number of reversals in four million year intervals as a histogram. The reversal frequency is relatively high in the interval 124-150 Ma, but appears to drop gradually to zero at the beginning of the so-called Cretaceous Normal Superchron (CNS), a period of some 40 m.y. in which no (or very few) reversals occurred. Since the end of the CNS at about 84 Ma, the frequency of reversals has increased to the present average rate of about four per million years.

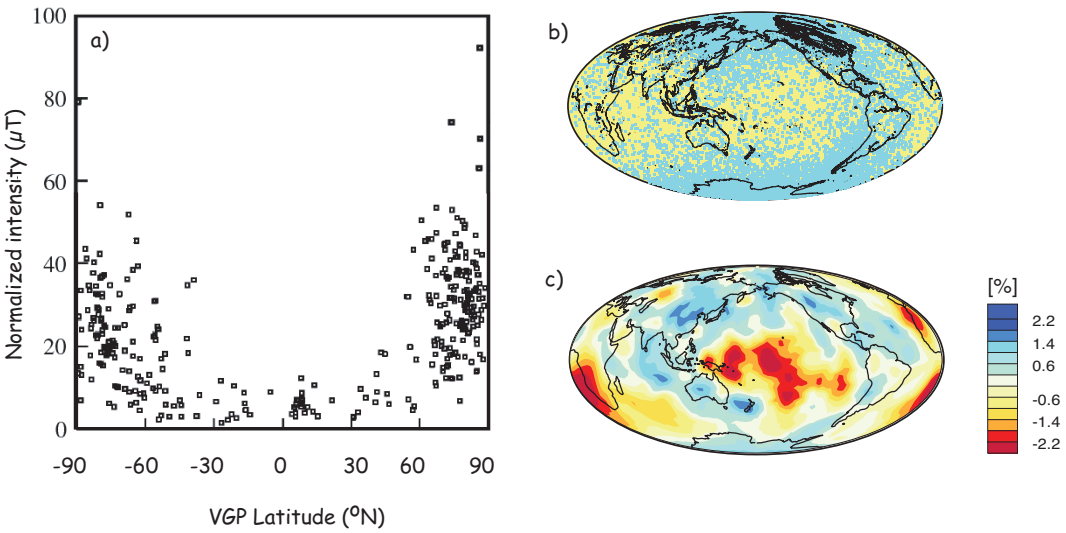


Figure 14.9: a) Normalized intensity (see Love, 2000) versus VGP latitude. As is seen in every data set of its kind, intermediate (transitional or excursional) directions (low VGP latitudes) are associated with low paleointensities. b) Plot of transitional VGPs from the TRANS data base. No selection criteria were applied. c) Shear wave velocity SB448 model from Masters et al. (2000) evaluated at 2770 km (core mantle boundary region). There is a fast (cold) ring around the Pacific, presumably from the influence of subducted slabs.

14.5 The time averaged field

It now seems possible to determine full scale spherical harmonic models as a function of time back thousands of years. Beyond a certain limit, however, there simply are not enough data with sufficient age control and spatial density to go back very far in time in this sort of fashion. The approach for longer time scales has been to look at the average magnetic field. There has therefore been a large effort over the past decade or so to bring together data for the purpose of evaluating the time averaged field.

The last five million years has been a recent focus because of the reduced effects of plate motion (e.g., Johnson and Constable, 1997). Data from lava flows from all over the world have been compiled into various databases and analyzed from a variety of view points. It was recently realized that the data had been compiled using less than optimum criteria and that many more data of higher overall quality may be required for a robust TAF model to be produced. Data from the new TAF project are only just becoming available. In the mean time, I show a plot of the TAF model of Hatakayama and Kono (2002) in Figure 14.11. Although the field is not perfectly GAD, the flux lobes seen in the historical field are nearly erased. The remaining non-dipole features may be artifacts of bad data.

It is worthwhile mentioning at this point that one of the primary assumptions in many paleomagnetic studies is that the magnetic field, when averaged over sufficient time, averages to that of a GAD field. This means that if VGPs are averaged from units spanning enough time to average out secular variation, the mean pole is coincident with the spin axis. Such a pole is called a *paleomagnetic pole*. As continents move, they carry with them rock units that retain a record of the spin axis

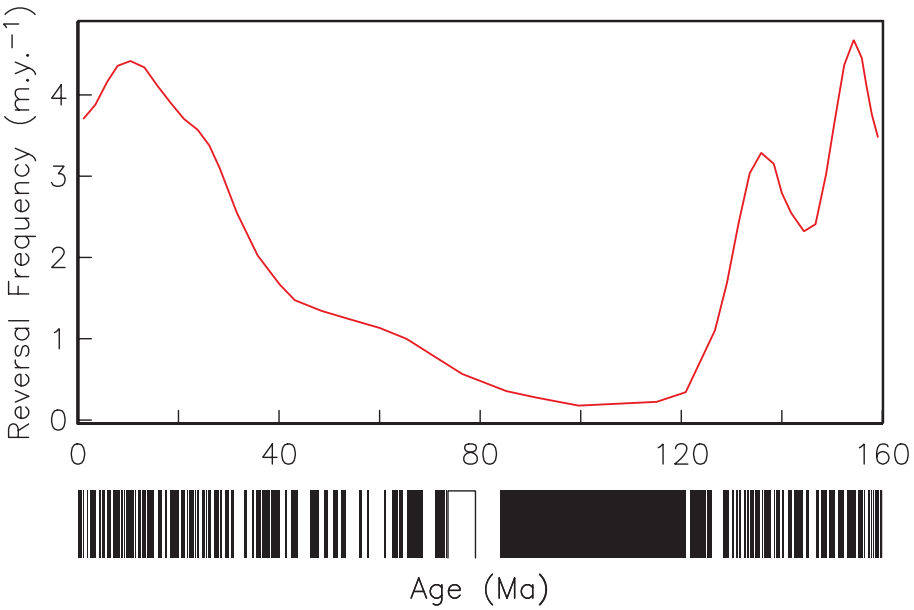


Figure 14.10: Barcode: The Geomagnetic Polarity Time Scale (GPTS) for the last 150 Ma (Berggren et al., 1995; Gradstein et al., 1995). Line traces the reversal frequency (number of reversals in a four million year interval) estimated by Constable (2003).

in the continental reference frame, so these poles tend to form swaths called *apparent polar wander paths* or APWPs. We will learn more about APWPs in later lectures. What is interesting to note now is that how much time is required to average out secular variation is not really known, but is more than 400 years and less than 5 million. Most text books claim that 10^4 - 10^5 years is sufficient (e.g., Butler, 1992). Also poorly defined is the minimum number of sampling sites required for a "good" average. Butler (1992) recommend at least ten, while ranging from ten while (Tauxe et al., 2003) suggest that approximately 100 sites are required to fully sample secular variation.

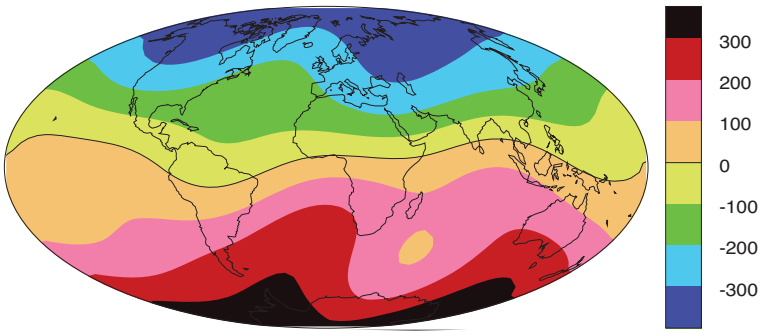


Figure 14.11: Time averaged radial magnetic field evaluated at the core mantle boundary. [Data from Hatakeyama and Kono. 2002.]

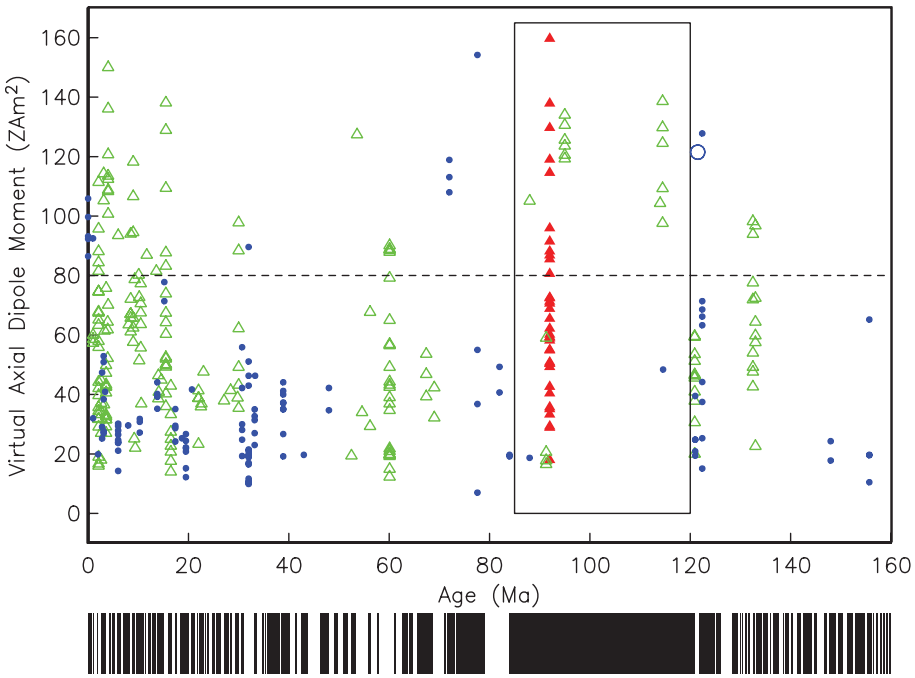


Figure 14.12: Summary of VADM data versus age compiled by Tauxe (in press). Green triangles are from lava flows. Red triangles arebasaltic glass data from the Troodos ophiolite. Dots are the DSDP/ODP submarine basaltic glasses. The present field is the dotted line. The thin solid line is the average dipole moment calculated assuming a log-normal distribution of the glass data. At the bottom is the Geomagnetic Polarity Time scale. [Figure from Tauxe, in press.]

14.6 Long term changes in paleointensity

As well as retaining a record of the direction of the magnetic field when cooled from high temperature,the magnetization in a rock has an intensity that is also a function of field magnitude. It is sometimes possible to estimate the magnitude of the Earth's field from geological samples. We plot a compilation of such data since the Jurassic in Figure 14.12, from Tauxe (in press). Early compilations suggested that much of the Mesozoic had a rather low field intensity (the *Mesozoic dipole low* of Prévot et al. (1990) with an apparent average intensity of about 25% of the present field which is ~ 80 ZAm^2. The compilation of high quality paleointensity data by Tauxe (in press) shows that the Cenozoic also had a predominantly low field, suggesting that the Mesozoic "dipole low" is probably the most common state of the geomagnetic field, with anomalously high values occurring in the latter part of the Cretaceous and early Cenozoic and during the last few thousand years.

14.7 Statistical models of paleosecular variation

From studies of the time averaged field it seems that, at least for the last five million years, the field has been dominantly that of a geocentric axial dipole (GAD). At any particular instant in time, however, there will be significant deviations owing to the non-axial dipole contributions. This,

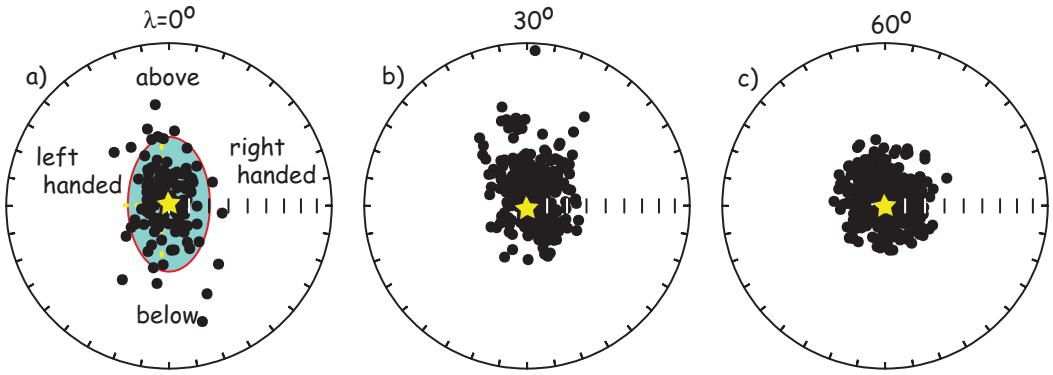


Figure 14.13: a) Paleomagnetic directions from the PSVRL database (see McElhinny and McFadden, 1997) compiled for latitude band 0-5° (N&S). Antipodes of reverse directions are used. The expected direction is at the star at the center of the equal area projection. Directions in the upper (lower) half are above (below) those expected and those to the right (left) are right-handed (left-handed). The red ellipse illustrates the elongation E of the directional data where E is the ratio of the eigenvalues along the maximum and minimum axes (here vertical and E-W respectively). b) Same as a) but for 25-35° (N&S) latitude band. c) Same as a) but for 55-65° (N&S) latitude band.

combined with distortions in the recording process (some of which were discussed in Lecture 5) and decreasing preservation of rocks with increasing age makes evaluating the GAD hypothesis increasingly difficult as we go back in time.

There has been considerable effort in collecting the data relevant to describing the statistical character of the geomagnetic field over time. Selected results from one such collection (that of McElhinny and McFadden, 1997; MM97) are shown in Figure 14.13. Directions from lava flows less than five million years old from particular latitudinal bands are plotted with respect to the expected GAD direction at that particular latitude (D', I' from Lecture 2).

Several things are worth mentioning about the data in Figure 14.13. First, it appears that the equatorial data are more elongate than those from higher latitudes (something we mentioned in Lecture 11). The elongation parameter E can be used to quantify this and is the τ_2/τ_3 ratio where τ_i are the eigenvalues of the orientation matrix (see Appendix C of Lecture 9). Secondly, the scatter in the directional data seems to go down with increasing latitude. The MM97 database culls data with VGP latitudes at an arbitrary angle away from the poles which results in a rather peculiar distribution of directions for the high latitude sites (Figure 14.13c). Third, when the directions are converted to VGP latitudes the scatter increases with increasing latitude.

VGP scatter is quantified by the parameter S (e.g., Cox 1969), defined as:

$$S^2 = (N-1)^{-1} \sum_{i=1}^{N} (\Delta_i)^2$$

where N is the number of observations and Δ is the angle between the i^{th} VGP and the spin axis. S' is the scatter calculated using the VGP collections that were culled of low latitude VGPs and this parameter is shown for the MM97 dataset in Figure 14.13d.

Most early modeling efforts by the paleomagnetic community went toward explaining the scatter in VGPs with latitude. This is but one of the many interesting and useful observations about the

statisitical behavior of the magnetic field and it would be wonderful if we had a way of predicting for a given latitude the full vector distributions expected from the geomagnetic field. To find a "full service" statistical paleosecular variation model, we begin with the work of Constable and Parker; hereafter CP88).

The CP88 statistical paleosecular variation model assumes that the time varying geomagnetic field acts as "Giant Gaussian Process" (GGP) whereby the gauss coefficients (see Lecture 2) g_l^m, h_l^m (except for the axial dipolar term, g_1^0 and in some models also the axial quadrupole term g_2^0) have zero mean. The standard deviations (see Figure 14.14a) are a function of degree l and a fitted parameter α (as in Figure 14.14b), and follow the formula:

$$\sigma_l^2 = \frac{(c/a)^2 l \alpha^2}{(l+1)(2l+1)} \tag{14.1}$$

where c/a is the ratio of the core radius to that of Earth (0.57). Many data sets show a persistent offset in equatorial inclinations at least in reverse polarity data sets, consistent with a small non-zero mean axial quadrupolar term ($\bar{g}_2^0$). We are ignoring this effect here because it is in all studies a small term.

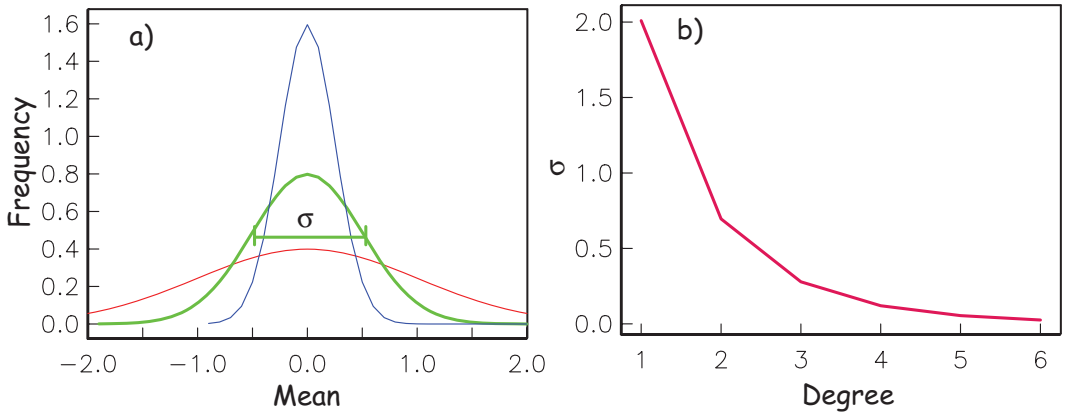


Figure 14.14: a) Illustration of a normal distribution with varying standard deviations. b) Variation of standard deviation σ as a function of spherical harmonic degree l in the CP88 model.

Once the average dipole moment $\bar{g}_1^0$, its standard deviation σ_1^0 and α are fixed, realizations of field models can be created by drawing the gauss coefficients from their respective gaussian distributions. Geomagnetic vectors can then be calculated for any given location using the usual transformation from the geomagnetic potential equation to geomagnetic elements (see Lecture 2).

The principal drawback of the CP88 model is that it fails to fit the observed scatter in the paleomagnetic data with latitude. Most of the subsequent variations on this theme attempted to address the VGP scatter problem by introducing more fitted parameters, losing the elegant simplicity of the CP88 model.

The most recent model of the statistical paleosecular variation genre is the TK03.GAD model of Tauxe and Kent (2004). Like CP88, TK03.GAD has only three parameters: $\bar{g}_1^0$ (set to fit a recent estimate for the long term average intensity of the axial dipole as in Figure 14.12), α as defined in CP88, but fit to the more recent compilation of directional data of McElhinny and McFadden (1997) and a new paramter β which is the ratio of the asymmetric ($l+m$ odd) to the symmetric ($l+m$ even) gauss coefficients for a given l. We show the variation in σ with degree for

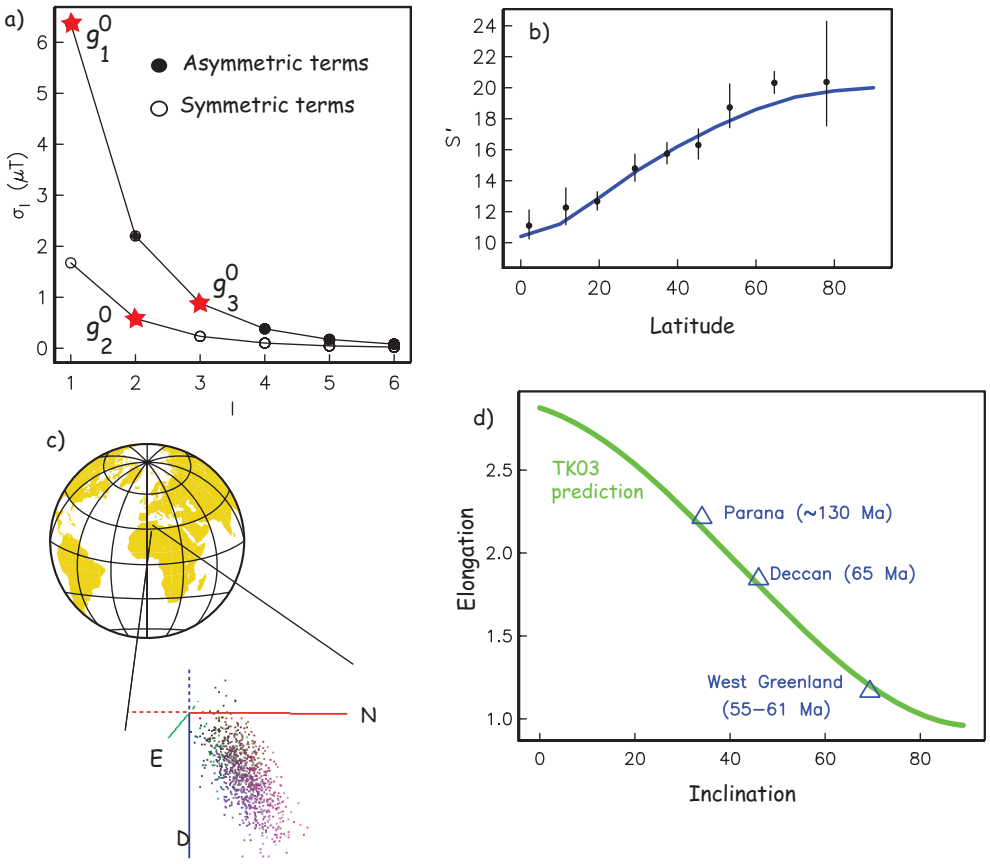


Figure 14.15: a) Variation of the standard deviation σ_l as a function of harmonic degree l for asymmetric and symmetric terms for the statistical field model TK03.GAD. All terms have zero mean except the axial dipole term. b) Estimated behavior of S' from the data compilation of McElhinny and McFadden (1997) (circles). Blue line is the predicted variation of S' from the TK03.GAD model of Tauxe and Kent (2004). c) 1000 vector endpoints from realizations of model TK03.GAD at 30°N. d) Elongation versus inclination predicted from the TK03.GAD model. Triangles are data from large igneous provinces back through time.

the two families (asymmetric and symmetric) in Fig. 14.15a. The term β allows a much improved fit to the paleomagnetic observations while the model retains the simplicity of the CP88 model (see Figure 14.15b).

In Fig. 14.15c we show the vector end points calculated from 1000 realizations of the model at 30°N. The distribution of these vectors predicts what would be observed at that latitude if we had a large number of observations of the geomagnetic field or its paleomagnetic proxies.

Models like TK03 can predict the distribution of geomagnetic field vectors at any location. These, then, can be compared with the observed paleomagnetic data in order to assess whether the data are consistent with the field model. The TK03 model was designed to predict values for S in agreement with those observed in the PSVRL database (see Figure 14.15b), but there are other attributes of the field that can be predicted as well. For example, while inclination can be calculated from the simple dipole formula (see Lecture 2) for any latitude, the elongation of the directions

(e.g., Figure 14.13a) requires a statistical field model. In fact, because elongation goes down with increasing latitude, while inclination goes up, there is a unique elongation inclination pair that is consistent with a given statistical field model. The elongation/inclination trend calculated from the TK03.GAD model is shown in Figure 14.15d.

Data from the last five million years fit the model predictions as it was designed to do, but the model can be tested through time by calculating the elongation/inclination pair for data sets of any age. The requirements are that the data are referenced to paleo-horizontal, that the directions represent the ancient geomagnetic field (they are not biased by overprinting, inclination error, etc.), and that there be a sufficient number to represent the statistical variability of the ancient geomagnetic field. There are not many data sets that satisfy these requirements. Three data sets from ancient large igneous provinces, however do: the Deccan Traps in India (Vandamme et al., 1991), the West Greenland volcanic province (Riisager et al., 2003) and the Paraná Basalts (Ernesto et al., 1999). The elongation/inclination pairs from these three data sets are plotted on Figure 14.15d for comparison with the model predictions. It appears that the TK03.GAD model can be used as a guide to the geomagnetic field behavior for at least the last 130 million years.

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