THE PALEOMAGNETISM OF SINGLE SILICATE CRYSTALS: RECORDING GEOMAGNETIC FIELD STRENGTH DURING MIXED POLARITY INTERVALS, SUPERCHRONS, AND INNER CORE GROWTH

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Received 6 November 2005; accepted 1 March 2006; published 31 March 2006.

[1] The basic features of the geomagnetic reversal chronology of the last 160 million years are well established. The relationship between this history and other features of the field, however, has been elusive. The determination of past field strength (paleointensity) is especially challenging. Commonly accepted results have come from analyses of bulk samples of lava. Historic lavas have been shown to faithfully record the past field strength when analyzed using the Thellier double-heating method. Data from older lavas, however, tend to show effects of in situ and laboratory-induced alteration. Here we review an alternative approach. Single plagioclase crystals can contain minute magnetic inclusions, 50–350 nm in size, that are potential high-fidelity field recorders. Thellier experiments using plagioclase feldspars from an historic lava on Hawaii provide a benchmark for the method. Rock magnetic data from older lavas indicate that the feldspars are less susceptible to experimental alteration than bulk samples. This resistance is likely related to the lack of clays. In addition, magnetic minerals are sheltered by the encasing silicate matrix from natural alteration that can otherwise transform the well-defined thermoremanent magnetization into an irresolute chemical remanent magnetization. If there is a relationship between geomagnetic reversal frequency and paleointensity, it should be best expressed during superchrons, intervals with few (or no) reversals. Thellier data sets based on single plagioclase crystals from lavas erupted during the Cretaceous Normal Polarity Superchron ($\sim$83–120 million years ago) suggest a strong ($>12 \times 10^{22}$ Am^2), stable field, consistent with an inverse relationship between reversal frequency and paleointensity. Superchrons may represent times when the pattern of core-mantle boundary heat flux allows the geodynamo to operate at peak efficiency, as suggested in some numerical models. Thellier data from single plagioclase crystals formed during times of moderate (<1 reversal/million years) and very rapid (>10 reversals/million years) reversal occurrence suggest a weaker and more variable field. These paleointensity data, together with a consideration of paleomagnetic directions, suggest that geomagnetic reversals, field morphology, secular variation, and intensity are related. The linkages over tens of millions of years imply a lower mantle control on the geodynamo. On even longer timescales the magnetization held by plagioclase and other silicate crystals can be used to investigate the Proterozoic and Archean geomagnetic field during the onset of growth of the solid inner core. Data from plagioclase crystals separated from mafic dikes, together with directional data from whole rocks, indicate a dipole-dominated field similar to that of the modern, 2.5–2.7 billion years ago. Older Archean rocks are of great interest for paleomagnetic and paleointensity investigations because they may record a time when the compositionally driven convection of the modern dynamo may not have been operating and a solid inner core did not play its current role in controlling the geometry of outer core flow. Most rocks of this age have been affected by low-grade metamorphism; investigations using single silicate grains provide arguably our best hope of seeing through secondary geologic events and reading the early history of the geodynamo. Absolute paleointensity measurements of the oldest rocks on the planet will require the further development of methods to investigate silicate crystals with exsolved magnetic minerals that address the uncertainties posed by thermocrystallization remanent magnetization, anisotropy, and slow cooling. Fortunately, prior work in rock magnetism, together with advances in analytical equipment and techniques, provides a solid foundation from which these frontier issues can be approached.

Citation: Tarduno, J. A., R. D. Cottrell, and A. V. Smirnov (2006), The paleomagnetism of single silicate crystals: Recording geomagnetic field strength during mixed polarity intervals, superchrons, and inner core growth, *Rev. Geophys.*, *44*, RG1002, doi:10.1029/2005RG000189.

1. MOTIVATION

[2] The Thellier double-heating method [*Thellier and Thellier*, 1959] and its common variant proposed by *Coe* [1967] are arguably the most rigorous means of learning about ancient field strength. In a typical experiment the natural remanent magnetization (NRM) of an igneous

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8755-1209/06/2005RG000189$15.00

Paper number 2005RG000189

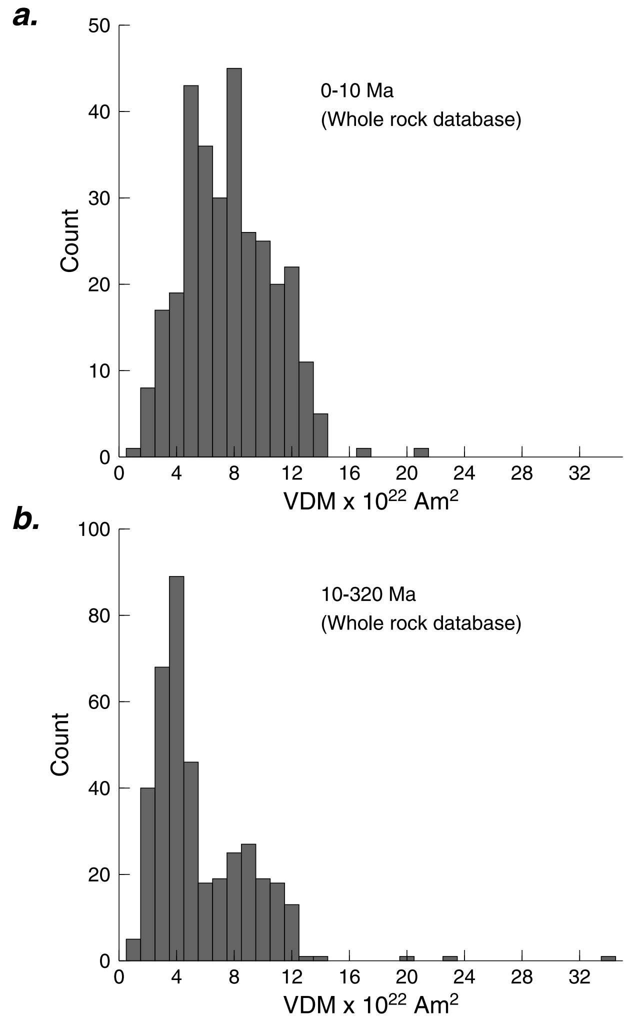

Figure 1. Histograms of virtual dipole moments derived from Thellier analyses of whole rocks formed between (a) 0 and 10 Ma and (b) 10 and 320 Ma [after *Tarduno and Smirnov*, 2004].

sample is demagnetized by heating to a given temperature and cooling in a field-free space. The sample is next reheated to the same temperature, but heating and cooling are done in the presence of a magnetic field. This "field-on" step results in the acquisition of a partial thermoremanent magnetization (pTRM). These paired "field-off" and field-on steps are repeated for increments that span the entire unblocking temperature range of a rock's magnetic minerals. When the pTRMs are additive, knowledge of the NRM lost (M_{NRM}), the thermoremanent magnetization (TRM) gained (M_{TRM}), and the applied field strength (H_{lab}) allows a calculation of the past field intensity (H_{paleo}):

$$H_{\mathrm{paleo}} = \frac{M_{\mathrm{NRM}}}{M_{\mathrm{TRM}}} H_{\mathrm{lab}}. \tag{1}$$

The additivity of pTRMs that lies at the heart of the Thellier method requires the magnetic recorders to be single domain, or single-domain-like, a state that is commonly called "pseudosingle" domain [*Stacey and Banerjee*, 1974]. After this requirement has been met, the next demanding aspect of paleointensity experiments centers on the field-on step; samples must not magnetically alter during heating. If alteration occurs, the TRM data will be corrupted. This is the factor that separates paleointensity studies from standard directional analyses where samples are demagnetized in a field-free environment and are thus less prone to the effects of experimentally induced alteration.

[3] Regardless of the experimental challenges, some authors have suggested that the mean field strength for the last 300 million years is relatively well known based on Thellier data derived from lava flows. However, is the venerable Thellier approach really ideal when applied to igneous whole rock samples? In tests using historic lavas, the method clearly works well because it retrieves the known field strength. Diligent efforts to detect experimental alteration and nonideal behavior related to magnetic mineral domain state characterize recent studies of lavas formed during the last 5 million years [e.g., *Laj and Kissel*, 1999; *Valet*, 2003]. However, in still older rocks the affects of weathering are omnipresent. Clay minerals begin to form a progressively larger portion of a whole rock's matrix. Magnetic mineral phenocrysts undergo low-temperature oxidation.

[4] During the successive heating steps required by the Thellier method, fine-grained magnetic minerals can form from clays [*Cottrell and Tarduno*, 2000]. Such alteration begins innocuously, with only slight changes in magnetic properties. At higher temperature increments and longer total times of heating the alteration can be obvious; it is easily detected by the now standard experimental methods [*Coe et al.*, 1978], when a pTRM is imparted at a lower temperature to check for the growth of magnetic minerals. The initial subtle stages of alteration in the laboratory, however, can elude detection.

[5] Low-temperature oxidation, on the other hand, results in maghemitization and a fundamental change in the nature of the magnetization [e.g., *O'Reilly*, 1984; *Özdemir and Dunlop*, 1985]. Rather than a TRM, the basis of the Thellier approach, the magnetization can become a chemical or crystallization remanent magnetization (CRM). The accuracy of CRM in preserving the original geomagnetic field strength is unclear [*Dunlop and Özdemir*, 1997].

[6] There are other signposts that should raise concerns about paleointensity estimates derived from older rocks. Principal among these is a difference in the mean field value for rocks formed before and after 10 Ma. In fact, many of the virtual dipole moments based on Thellier data from whole rock samples 10 to 320 million years old are comparable to values that characterize younger geomagnetic excursions and reversal transitions (<4 × 10^{22} Am^2 [*Chauvin et al.*, 1989; *Tarduno and Smirnov*, 2004]). This is paradoxical because trends in secular variation, as well as the presence of long intervals without geomagnetic reversals, imply stability of the pre-10 Ma geomagnetic field. Taken at face value, these Thellier data imply that the field has been extraordinarily energetic during the last 10 million years (Figure 1). A change in heat flux at the core-mantle

boundary induced by large-scale mantle flow [e.g., *Glatzmaier et al.*, 1999; *Olson and Christensen*, 2002; *Christensen and Olson*, 2003] could affect the energetics of the dynamo. However, this would require many tens of millions of years.

[7] Natural and related experimental alteration of lavas tends to lower paleointensity values derived from Thellier analyses [*Tarduno and Smirnov*, 2004]. Therefore the apparent difference in field strength before and after 10 Ma may be an artifact. This possibility evokes a much older debate. *Koenigsberger* [1938a, 1938b] noted systematic changes in remanence with age, favoring a decay in NRM [see also *Nagata*, 1953]. Thellier instead called for a change in magnetic field strength with time [*Thellier and Thellier*, 1959]. Today we adopt the Thellier experimental approach, but we also have a more comprehensive set of analytical tools that tell us that whole rocks are not impervious to change with time. This provides motivation for a search for natural paleomagnetic recorders that are less susceptible to alteration in nature and in the laboratory. Single silicate crystals can provide such an alternative recorder of geomagnetic field history.

2. EARLY STUDIES

[8] Studies of the magnetic behavior of rock-forming silicate minerals can be traced to early attempts to understand the magnetization of whole rock samples then being used to define the first apparent polar wander paths [*Creer et al.*, 1954; *Hospers*, 1955; *Runcorn*, 1956; *Irving*, 1956, 1964]. For some of these rocks the source of the magnetization was not immediately apparent. Phenocrysts of magnetic minerals could not be found using standard optical microscopes. It was accepted that silicate minerals were not ferrimagnetic or antiferromagnetic [*Nagata*, 1961]. However, inclusions of magnetic minerals had been observed in some silicate grains [e.g., *Bown and Gay*, 1959]. In clinopyroxene, two sets of inclusions, subparallel to the [100] and [001] crystallographic directions were noted. Ted (M. E.) Evans demonstrated that such inclusions could be important remanence carriers. In a series of papers published between 1966 and 1977, Evans and his coauthors demonstrated that the stable remanence recorded by whole rock samples of the Modipe gabbro of southern Africa was related to clinopyroxene grains containing exsolved magnetite [*Evans and McElhinny*, 1966; *Evans et al.*, 1968; *Evans and Wyman*, 1970; *Evans*, 1977]. Following this work, *Murthy et al.* [1971, 1981] described "needles" of single-domain (SD) or near-SD character oriented along the crystallographic planes of pyroxenes of the Michikamau anorthosite (Labrador).

[9] Exsolution in plagioclase was also recognized as interest in paleomagnetism grew. In some cases this exsolution was linked to the clouded appearance of plagioclase seen in some igneous rocks [*Poldervaart and Gilkey*, 1954; *Armbrustmacher and Banks*, 1974]. *Anderson* [1966] noted that Fe oxides may exsolve from structural positions within the crystal lattice of plagioclase and redeposit along twinning planes. *Hargraves and Young* [1969] separated oxide, plagioclase, and pyroxene from the Lambertville diabase and found that the silicates had the potential to make substantial contributions to the remanence. The presence of magnetite rods in plagioclase from oceanic gabbros was documented sometime later [*Davis*, 1981].

[10] Of particular note is the work done in Michael Fuller's group in the 1970s. Through an examination of oriented grains from the Tatoosh granodiorite it was found that plagioclase carried a stable remanence, while hornblende and biotite seemed to carry unstable multidomain magnetic grains [*Wu et al.*, 1974]. Given the instrumentation limitations at the time, this was a significant accomplishment. Other important studies include further documentation through electron microscopy of magnetic particles in silicates [*Morgan and Smith*, 1981; *Scofield and Roggenthen*, 1986; *Geissman et al.*, 1988; *Schlinger and Veblen*, 1989; *Harlan et al.*, 1994; *Xu et al.*, 1997]. The domain structures of magnetite inclusions in biotite and hornblende were also studied by *Özdemir and Dunlop* [1992] and were reported in more detail by *Dunlop and Özdemir* [1997].

3. ON CHOOSING SILICATES WELL

[11] Given the likely presence of fine-grained magnetic particles in single silicate minerals, which rocks should be chosen for analysis? Choosing those rocks with the largest crystals that could most easily be removed from the matrix is one approach. Such crystals might also yield the largest initial magnetizations. However, our selection must be guided by magnetization process rather than the ease of separation or measurement.

[12] The magnetization process of choice is a TRM acquired during cooling of a molten rock; we have the best developed theory for this process [*Néel*, 1949, 1955; *Dunlop and Özdemir*, 1997]. Furthermore, the chances that a straightforward TRM is imparted are enhanced when the temperature of the magma is well above the temperature at which the magnetic particles in the silicate crystals acquire a remanent magnetization. For example, the temperatures of typical basaltic magmas range from approximately 1200°C to 1000°C. There is a large difference between these temperatures and the typical upper bound on Curie temperature for a primary magnetic inclusion (~580°C for magnetite). For felsic magmas, which range in temperature between 800°C and 500°C, the gap narrows or disappears altogether. Interflow and intraflow motion of felsic lavas at temperatures close to the Curie points of magnetic inclusions is an important possibility. Some silicate crystals could acquire one, or several, partial thermoremanent magnetizations during solidification of the flow.

[13] Temperature history is also important. In most of the early studies discussed in section 2, the magnetic particles observed reflected exsolution during slow cooling. If this exsolution occurs above the Curie temperature of the magnetic mineral of interest (usually magnetite), the magnetization is a TRM. We can proceed with studies of magnetic directions. However, for paleointensity investiga-

tions we must first address the anisotropy that accompanies the formation of magnetic needles along crystallographic axes. Paleointensity corrections are needed to account for the potentially large differences between natural and laboratory cooling rates; hence independent information is needed on the rate of cooling. The sense of these corrections appears to depend on magnetic mineral domain state [*Halgedahl et al.*, 1980; *McClelland-Brown*, 1984], highlighting the need for good rock magnetic controls prior to paleointensity investigation.

[14] If exsolution continues below 580°C, the situation is much more complex [e.g., *Smirnov and Tarduno*, 2005]. The magnetization is at least in part a thermocrystallization remanent magnetization (TCRM), acquired during crystal growth. TCRM should be less efficient than TRM [*Dunlop and Özdemir*, 1997], but the quantification of the process is difficult. At the very least it requires even greater controls on magnetic mineral properties and cooling histories, and this information is often not readily available.

[15] Some magnetic particles may be trapped during the formation of a silicate grain and thus are not related to exsolution during slow cooling. Other silicate crystals may have exsolved magnetic particles, but they are xenocrysts that have clearly been incorporated into a rock at temperatures well above the Curie temperature of magnetite. Still other silicates might have inclusions exsolved at very high temperature, with a relatively small anisotropy. These possibilities optimize the magnetization process. The rock type that best encompasses these possibilities is the basaltic lava flow. Thin basaltic dikes might be expected to have similar characteristics.

[16] The principal drawback associated with silicates separated from these rocks is low natural remanent magnetization intensity, related at least in part to small crystal sizes (typically 1 mm or less). Until recently, the measurement of such small crystals was beyond the sensitivity of even RF superconducting quantum interference device (SQUID) magnetometers [*Goree and Fuller*, 1976]. The widespread availability of DC SQUIDS [*Kleiner et al.*, 2004], however, has resulted in significant increases in resolution. The measurement of these crystals has become feasible with smaller-bore instruments with high-resolution coil geometries.

[17] In section 4 we first review the progress made using plagioclase crystals from Mesozoic to recent basaltic lavas to study the geodynamo. Next we discuss the potential for using the magnetization of single silicate crystals for understanding the magnetic field of the young Earth. The latter discussion will require us to revisit the magnetization carried by exsolved magnetic particles in other rock types.

4. PLAGIOCLASE FELDSPARS FROM BASALTIC LAVAS

[18] To identify an optimal recorder, *Cottrell and Tarduno* [1997, 1998] surveyed the magnetic hysteresis properties of various silicate grain types separated from basalts. Olivine and pyroxene revealed pseudosingle to multidomain curves (saturation remanence with respect to saturation magnetization, M_r/M_s, values <0.10, and coercivity of remanence with respect to coercivity, H_{cr}/H_c, values $\geq$2.50), whereas plagioclase crystals often showed more single-domain-like behavior. This led to a focus on plagioclase crystals, which were generally handpicked from whole rock basalt samples after the latter were crushed with a ceramic mortar and pestle. The crystals are routinely washed and sonicated with distilled water. Crystals used in subsequent experiments typically range in size from 0.5 to 2.0 mm (longest dimension), are free of the rock groundmass, and are clear when viewed with an optical microscope.

[19] One of the nonintuitive aspects of this analysis is that crystals with large visible inclusions are avoided. If such inclusions are magnetic particles, the fact they can be seen by the naked eye (or under low-power magnifications) indicates that they will almost certainly be multidomain in nature and hence ill suited for Thellier analyses.

[20] Magnetic extracts are prepared by crushing plagioclase crystals to a fine power, suspending the powder in distilled water, and concentrating the magnetic particles with a rare earth magnet. Transmission electron microscopy (TEM) analyses of these magnetic extracts indicate that the plagioclase can contain equant to slightly elongated magnetic particles 50 to 350 nm in size [*Cottrell and Tarduno*, 1999, 2000] (Figure 2).

[21] The potential alignment of elongated magnetic particles within silicate grains is an important concern for Thellier paleointensity experiments. Magnetic particle alignment could affect the acquisition of pTRM because the easy direction of magnetization along an elongated particle should have a lower coercivity [e.g., *Dunlop and Özdemir*, 1997]. The anisotropy of magnetic hysteresis parameters is one indication of alignment of magnetic inclusions. To test for an anisotropy, plagioclase grains separated from basalts have been measured at various angles using a Princeton Measurements Corporation alternating gradient force magnetometer [*Flanders*, 1988] (Figure 2). Crystals oriented in planes both parallel and perpendicular to the applied field have been measured. Studies of this potential directional dependence of magnetic hysteresis indicate anisotropies in the plagioclase crystals ranging from 0 to 3%. The higher value could result in only a few degrees of pTRM deflection during a typical Thellier experiment [*Cottrell and Tarduno*, 1999, 2000; *Tarduno et al.*, 2002a].

[22] The unblocking temperatures of the plagioclase crystals were found to be similar to those of the whole rocks from which they are derived, implying a similarity of composition of the magnetic minerals. For basaltic rocks these unblocking temperatures were often most consistent with a titanomagnetite carrier. This is an important observation because the composition of exsolved magnetic particles would be expected to be more similar to end-member phases.

[23] Thermal demagnetization through warming of a saturation isothermal remanent magnetization acquired at 20 K has shown the presence of a blurred Verwey

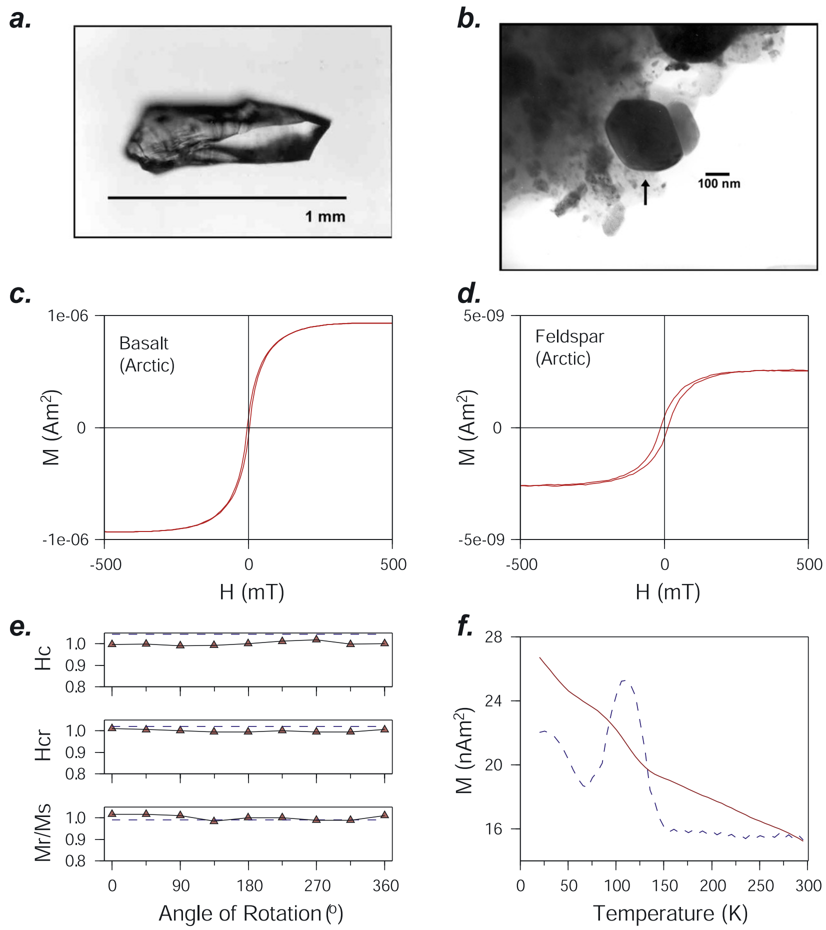

Figure 2. Optical, transmission electron microscope (TEM), and rock magnetic analyses of whole rocks, plagioclase crystals, and magnetic separates. (a) Typical plagioclase crystal used for rock magnetic and paleointensity experiments. (b) TEM image of a magnetic separate from a plagioclase crystal. Magnetic hysteresis data (slope corrected) for a whole rock (c) basalt sample and (d) plagioclase feldspar. In general, plagioclase crystals show single-domain to pseudosingle-domain behavior, while whole rocks display pseudosingle to multidomain characteristics. The former are better suited for Thellier paleointensity analyses. (e) Normalized hysteresis parameters versus rotation angle measured on a plagioclase crystal. The crystal was rotated by ~45° increments on the stage of a Princeton Measurements Alternating Gradient Force Magnetometer parallel (solid line and triangles) and perpendicular (dashed line, only mean shown) P1 probes. Abbreviations are M_r/M_s, saturation remanence/saturation magnetization ratio; H_{cr} coercivity of remanence; and H_c, coercivity. The lack of systematic variations of these parameters indicates that anisotropy is not a significant concern for Thellier paleointensity experiments. (f) Warming curve of a magnetization (solid line) acquired by a plagioclase crystal at 20 K in a 2.5 T field. Dotted line is the inverse of the derivative. Examples are from the Strand Fiord basalts of the high Arctic after *Tarduno et al.* [2002a].

transition (the transition from cubic to monoclinic crystalline symmetry [*Verwey*, 1939; *Moskowitz et al.*, 1998]) in plagioclase crystals separated from Cretaceous basalts of the Arctic [*Tarduno et al.*, 2002a] (Figure 2). These data again suggest that the plagioclase crystals contain magnetic inclusions of composition similar to those of magnetic grains in the whole rock. As shown by the low-temperature data, the carrier is likely a low-Ti titanomagnetite. This observation further suggests that the magnetic particles are inclusions incorporated during plagioclase crystallization [*Smith*, 1975].

4.1. Paleointensity Methods

[24] In general, cleaned plagioclase with NRM intensities $>\sim 5 \times 10^{-11}$ Am^2 have been selected for paleointensity measurements. For early studies, crystals were set inside standard salt pellets [e.g., *Tarduno and Kodama*, 1982], 0.5 inch in diameter, for paleointensity analysis. The relatively large size of these pellets aided orientation during the Thellier experiment. It also allowed, in many cases, recovery of the crystal after the experiment so that magnetic hysteresis properties could be measured to check for experimental alteration. The disadvantage of the size of these

pellets is the increased chance of contamination and the related increase in the magnetization of the blank. To address this issue, small quartz tubes (3 mm internal diameter) have been used in later studies [*Tarduno et al.*, 2002a]. After the tubes are cleaned in acid and their magnetization checked with a SQUID magnetometer, the crystals are set inside the tubes with Omega cement (which itself is checked for contamination). A very small quartz tube thus becomes the sample holder, reducing greatly the nonsample mass that is positioned within the magnetometer sensing region during measurement.

[25] Applied fields (40 or 60 μT) in an ASC Scientific thermal demagnetization oven have been used for the modified Thellier experiment. No significant differences in paleointensity values have been noted in experiments using these field values. In general, temperature steps of 50°C have been used for the first 250°C (an unblocking temperature range dominated by viscous overprints in many basaltic samples), followed by demagnetization and pTRM acquisition at 25°C increments until >90% of the natural remanent magnetization was lost or the magnetizations were no longer stable. For some samples, smaller temperature increments (e.g., 10°C–15°C) have been used after treatment at 250°C to better constrain the ideal temperature range for analysis. Orthogonal vector plots of "field-off" steps are analyzed to evaluate the temperature range of potential secondary magnetizations and to further determine the temperature range with optimal directional consistency to be examined in NRM/TRM plots.

[26] Several criteria have been used to judge the data. Successful analyses must show a linear relationship between NRM lost and TRM gained, where the slope of the line is the ratio of the ancient field to the applied field. Least squares analysis is used to fit the data. Four or more points must define the best fit line, with R^2 values >0.90. A significant percent of the primary component remanence (typically greater than 50%) should be lost within the temperature range defined by the best fit line. Ideally, NRM/TRM points should be evenly distributed along the best fit line, although exceptions are sometimes warranted in the case of magnetic carriers with very narrow unblocking spectra.

[27] Several pTRM checks are performed to check for alteration. To be judged successful (i.e., lack of alteration), a pTRM check must fall within 5% of the original TRM value. Least squares fit of the directional data of the field-off steps of the modified Thellier-Thellier technique [i.e., after *Coe*, 1967] must also have a mean angular dispersion of <15°and show a trend toward the origin in orthogonal vector plots. In addition, the inclination direction of the field-off steps should not tend toward the applied field (I = 90° for samples oriented with their z axis parallel to the direction of the applied field).

[28] Changes in NRM and TRM intensities during the Thellier experiments using plagioclase crystals are typically very small (sometimes $<5 \times 10^{-12}$ Am^2). Background measurements on the superconducting rock magnetometer used in these experiments (a 2G Enterprises DC SQUID magnetometer with a 4 cm diameter access and high-resolution sensing coils) range from 4 to 8×10^{-13} Am^2. A "blank" salt pellet can have intensities ranging from 5×10^{-12} to 1.5×10^{-11} Am^2. To reduce the influence of measurement noise, a three-point sliding average of NRM and TRM intensities was used over the optimal temperature range identified in the orthogonal vector plots. These averaged data were used in assessing the quality of the data through reliability checks and in the calculation of final paleointensities [*Cottrell and Tarduno*, 2000; *Tarduno et al.*, 2001]. Typical 3 mm quartz sample holders have magnetizations of about $1–2 \times 10^{-12}$ Am^2, significantly less than salt pellets. This reduction in the blank has led to less of a reliance on averaging [*Tarduno et al.*, 2002a]. In the most recent studies, only raw NRM/TRM values were used in assessing the data and calculating paleointensities [*Tarduno and Cottrell*, 2005].

4.2. Success Rates: There Are No Paleointensity Panaceas

[29] Success rates based on over 400 experiments to date using single plagioclase crystals range from about 33% to slightly over 50%. While these values compare well with the success rate of Thellier analyses of lava whole rock samples scrutinized with rigorous reliability criteria (generally a success rate of ~20%), it, nevertheless, testifies that the approach is not a quick, easy, or cheap means of obtaining past field information. Instead, the focus is on higher accuracy.

[30] Crystal results are rejected for a variety of reasons. Many crystals exhibit a sharp drop in NRM intensity after the first few demagnetization treatments. These crystals carry some type of secondary magnetization acquired either in nature or in the sample preparation process; the reason that they fail in the paleointensity analysis, however, is that after this initial intensity drop, further measurement becomes impractical. That is, the NRM intensity used to select the crystals for analysis was artificially high because of the overprint component. One simple way to help avoid this problem is to place crystals in a field-free space prior to selection for the Thellier experiment and to monitor viscous decay of potential overprints through NRM measurements with time.

[31] Other crystals do not show stable behavior (either in NRM decay or pTRM acquisition) throughout the experiment. A few percent of samples have field-off steps that do not tend to the origin, and a few show field-off directions that tend toward the direction of the applied field.

[32] It is likely that some crystals will not have a smoothly decaying remanence because of undesirable grain size distributions. However, for others this behavior may still be related to experimental noise given the low NRM intensities. Therefore further improvements in the practical sensitivity of three-component DC SQUID magnetometers could improve experimental success rates. In addition, the further development of other approaches, such as those using the low-temperature superconducting SQUID microscope [*Wikswo*, 1996; *Weiss et al.*, 2001] or the atomic

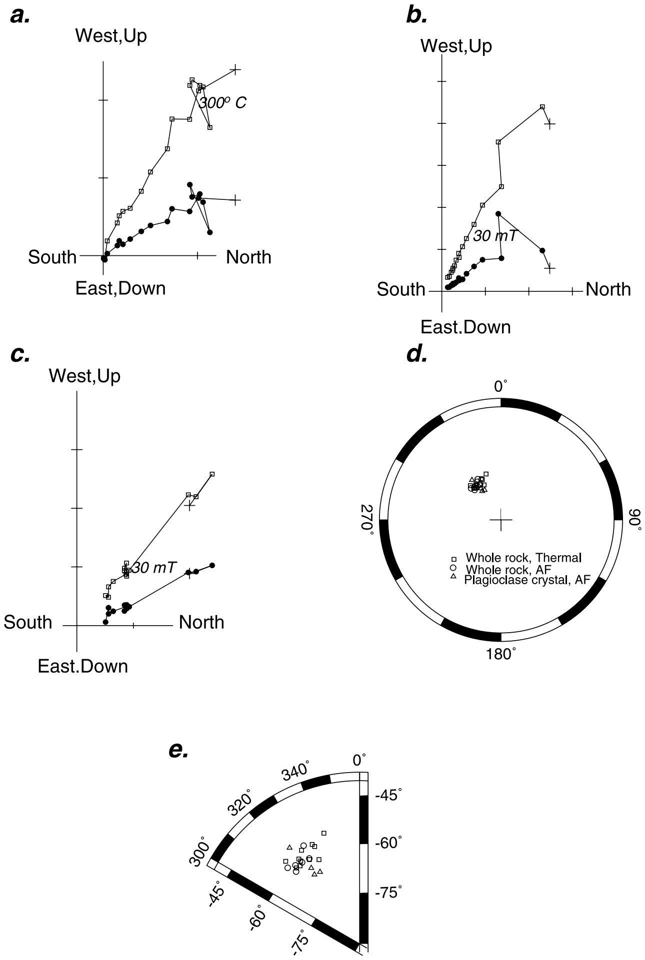

Figure 3. Paleomagnetic directions from oriented plagioclase grains compared to those from whole rocks [*Cottrell and Tarduno*, 2000]. (a) Orthogonal vector plot of whole rock sample treated with thermal demagnetization, with 25°C temperature steps. (b) Orthogonal vector plot of whole rock sample treated with alternating field demagnetization, with 5 mT increments to 80 mT. (c) Orthogonal vector plot of oriented plagioclase crystals (see text) treated with alternating field demagnetization, with 5 mT increments. (d) Stereographic projection of directions derived from thermal demagnetization of whole rocks, alternating field demagnetization of whole rocks, and alternating field demagnetization of oriented plagioclase crystals. (e) Expanded view of Figure 3d.

magnetometer [*Kominis et al.*, 2003; *Fitzgerald*, 2003], notwithstanding debate on the subject [e.g., *Wikswo*, 2004], might ultimately assist efforts to read the paleomagnetic signal of the weakest silicate crystals.

4.3. A Modern Field Benchmark

[33] To test the method, Thellier experiments were conducted on plagioclase separated from a 1955 flow from Kilauea, Hawaii [*Macdonald and Eaton*, 1964]. As the field is known from records at the Honolulu magnetic observation, this flow is ideal for testing paleointensity methods, and it has been the target of previous efforts toward that goal [*Doell and Smith*, 1969; *Coe and Grommé*, 1973]. The plagioclase in the flow studied are euhedral, indicating their presence in the magma chamber prior to eruption [*Wright and Fiske*, 1971]. A diffusive cooling model suggests that the ~1 m thick flow front cooled in hours to days. This further suggests that paleointensity corrections that are sometimes needed to account for the differences in experimental and natural cooling, corrections which become very important for slowly cooled intrusive rocks [e.g., *Halgedahl et al.*, 1980], are in this case negligible. The paleointensity estimates obtained from single plagioclase crystals agreed with values reported in detailed Thellier analyses of whole rocks [*Coe and Grommé*, 1973] and from the magnetic observatory data [*Cottrell and Tarduno*, 1999].

4.4. Do Plagioclase Crystals and Whole Rocks Record the Same Direction?

[34] To test whether plagioclase recorded the same direction as the bulk sample, *Cottrell and Tarduno* [2000] prepared samples of oriented crystals from a typical lava flow of the Rajmahal Traps. Through a combination of successive grindings and etchings to remove the basalt matrix (and nonplagioclase phenocrysts) an oriented thin section of plagioclase crystals was made. These thin sections were then trimmed to further isolate crystals that were optically clean, and the slide was sonicated to remove potential residue from the grinding process. Alternating field demagnetization (AF) of the thin sections (which contained 4–6 crystals) yielded data that agreed with AF and thermal demagnetization of the whole rocks (Figure 3).

4.5. Whole Rocks Versus Single Plagioclase Crystals

[35] Further comparisons of rock magnetic data from whole rock samples and plagioclase crystals taken from a Rajmahal lava flow demonstrated that the plagioclase crystals were less altered by Thellier heatings [*Cottrell and Tarduno*, 2000]. Specifically, magnetic hysteresis data from heated whole rock samples indicated the massive formation of a fine-grained magnetic phase; this behavior was not seen in the plagioclase crystals (Figure 4). This difference in rock magnetic behavior with heating paralleled differences seen in the fidelity and absolute values of Thellier paleointensity data. Although whole rock samples meeting paleointensity reliability criteria yielded paleointensity values that were within error of those derived from single plagioclase crystals, such samples were rare. In general, the whole rocks seldom met reliability criteria and yielded low nominal paleointensity values. The differences can be attributed to the formation of fine-grained magnetite from clays in the groundmass of whole rock samples, resulting in the anomalous acquisition of TRM during Thellier experiments and a bias toward low calculated paleofield values.

5. PALEOINTENSITY, SECULAR VARIATION, FIELD MORPHOLOGY, AND REVERSAL FREQUENCY

[36] The experimental alteration, nonideal behavior, and associated high rates of sample reject that typify the Thellier method as applied to whole rocks have generally resulted in

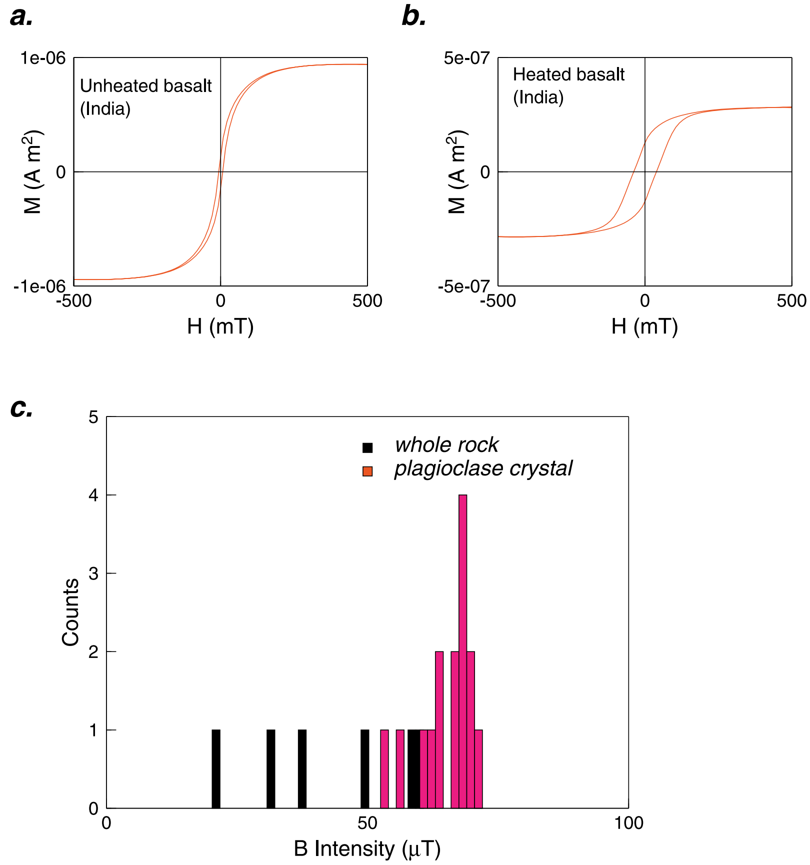

Figure 4. Analyses of whole rocks and plagioclase crystals from a lava flow of the Rajmahal Traps [after *Cottrell and Tarduno*, 2000]. Slope-corrected magnetic hysteresis curves for (a) unheated and (b) heated whole rock samples. The curve for the heated sample documents the growth of a fine-grained magnetic phase. The heating increments applied were those of a typical Thellier experiment. Magnetic hysteresis curves for plagioclase crystals did not show significant changes after heating (see discussion of *Cottrell and Tarduno* [2000]). (c) Comparison of paleointensity results from whole rock samples and plagioclase crystals.

a piecemeal approach to the definition of the past geomagnetic field. A directional study from one place and time is combined with a paleointensity value from another place and time to compose a picture of the field. The underlying spatial and temporal variation of intensity and directions are unknown. Given that the data are sparsely populated in time, the conclusions are fragile.

[37] The ability to derive paleointensity values from plagioclase crystals from long lava sequences provides the chance to look at all parts of the time-averaged field at a single location. The background chronology of field reversals for the last 160 million years is already well established from the record of seafloor marine magnetic anomalies and magnetostratigraphic studies [*Opdyke and Channell*, 1996]. To investigate field geometry, we consider the geomagnetic field at a radius r, colatitude θ, and longitude ϕ, which can be described by the gradient of the scalar potential (Φ):

$$\Phi(r,\theta,\phi) = r_e \sum_{l=1}^{\infty} \sum_{m=0}^{l} \left(\frac{r_e}{r}\right)^{l+1} P_l^m(\cos\theta)\left[g_l^m \cos m\phi + h_l^m \sin m\phi\right], \quad (2)$$

where P_l^m are partially normalized Schmidt functions and r_e is the radius of the Earth. The Gauss coefficients g_l^m and h_l^m describe the size of spatially varying fields. It is clear that the field for Mesozoic to recent times has been dominated by the axial dipole field (g_1^o), but there is continued debate on the potential contributions of higher-order terms such as the axial octupole (g_3^o) [e.g., *Van der Voo and Torsvik*, 2001; *Courtillot and Besse*, 2004]. Paleomagnetic directions from whole rock samples of lavas can be used to investigate this question, especially when the data are compared with other results from varying latitudes. Paleointensity data from plagioclase crystals separated from the same rocks can be used to define past field strength. However, given the considerable experimental demands of paleointensity studies using single crystals, examining the potential relationships between geomagnetic reversal frequency, secular variation, field morphology, and paleointensity is still a formidable task. If connections exist, however, they should be best expressed during superchrons, intervals tens of millions of years long with few (or no) reversals [e.g., *Jacobs*, 2001]. Hence the goal of examining all components of the geomagnetic field is approachable if one focuses on a superchron.

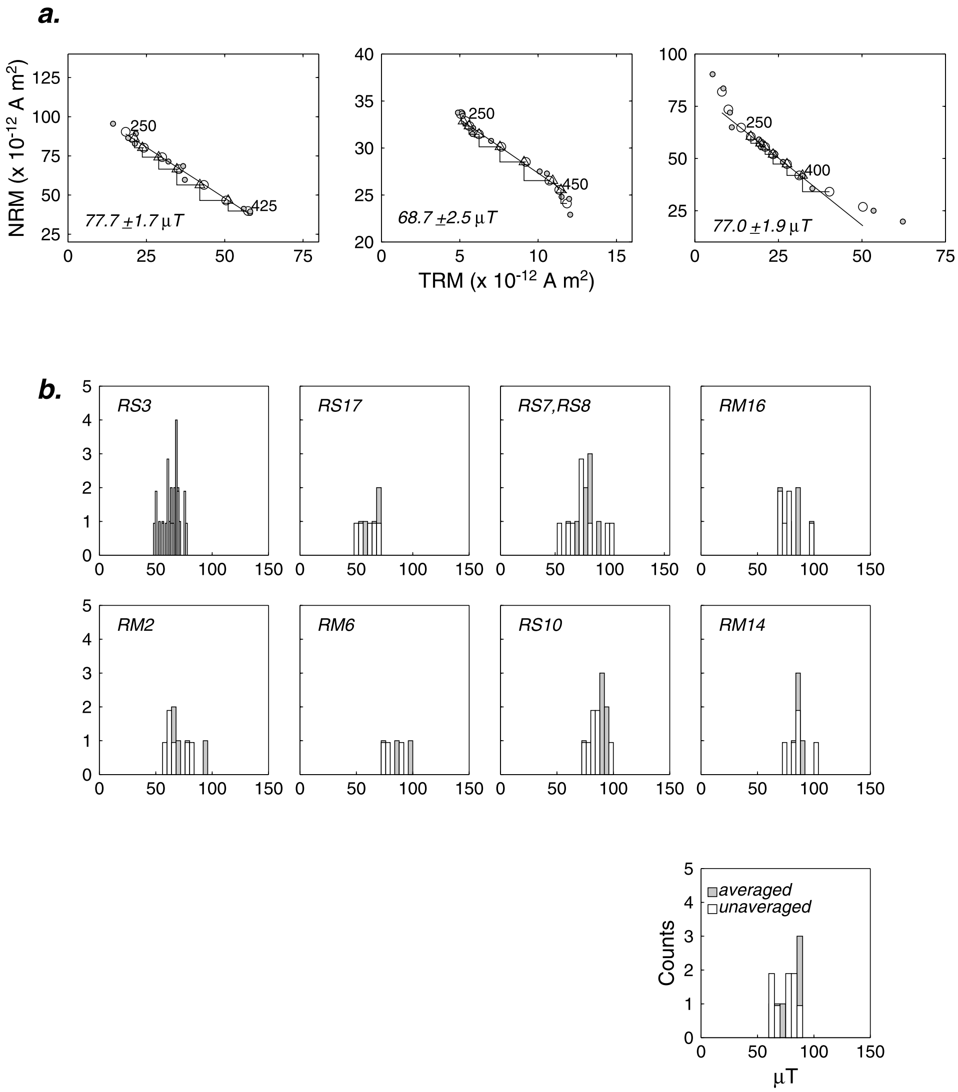

Figure 5. Paleointensity results from plagioclase crystals from the ~113–116 Ma Rajmahal Traps [after *Tarduno et al.*, 2001]. (a) Typical plots of natural remanent magnetization (NRM) versus thermal remanent magnetization (TRM). Partial thermoremanent magnetization (pTRM) checks are shown by triangles. Large open circles are three-point sliding-window averages of raw data (small circles). Labeled points are temperatures (°C) representing the range used to determine paleointensity. (b) Summary of paleointensity data meeting reliability criteria from the Rajmahal Traps. See *Tarduno et al.* [2001] for site locations (RS3, RS17, RS7, RS8, RM16, RM2, RM6, RS10, and RM14). Averaged data are shown as shaded; unaveraged data are shown as unshaded.

5.1. Paleointensity During the Cretaceous Normal Polarity Superchron

[38] The Rajmahal Traps, which were erupted at 113 to 116 Ma [*Kent et al.*, 1997] during the Cretaceous Normal Polarity Superchron, provide an opportunity to gain this more complete view of the past geomagnetic field [*Tarduno et al.*, 2001]. Thellier paleointensity analyses of single plagioclase have been derived from eight distinct lava units (here the term "lava unit" is used because it is sometimes necessary to combine results from adjacent lava flows that could have been erupted in a very short time). These units were selected from a larger data set so that they span secular variation. This was gauged in two ways. First, the stratigraphic position of the lava units was considered; units were selected to span the sequence and, in particular, sedimentary horizons. Second, the angular dispersion of paleomagnetic directions of the select units (derived from analyses of whole rock samples) mimicked that of a complete sampling of the lava sequence.

[39] Fifty-six of the 149 crystals measured met selection criteria (Figure 5). The averaged data yield a mean intensity of 77.6 ± 9.4 μT (1σ uncertainty). This value is not significantly different from that obtained from fitting the unaveraged data (74.5 ± 9.6 μT). Because the stratigraphic and angular dispersion data indicate that secular variation has been averaged, the paleointensity data allow the calculation of a time-averaged dipole strength known as a paleomagnetic dipole moment. The resulting value of $12.5 \pm 1.4 \times 10^{22}$ Am^2 [*Tarduno et al.*, 2001] is higher than the present-day field or prior estimates of the long-term

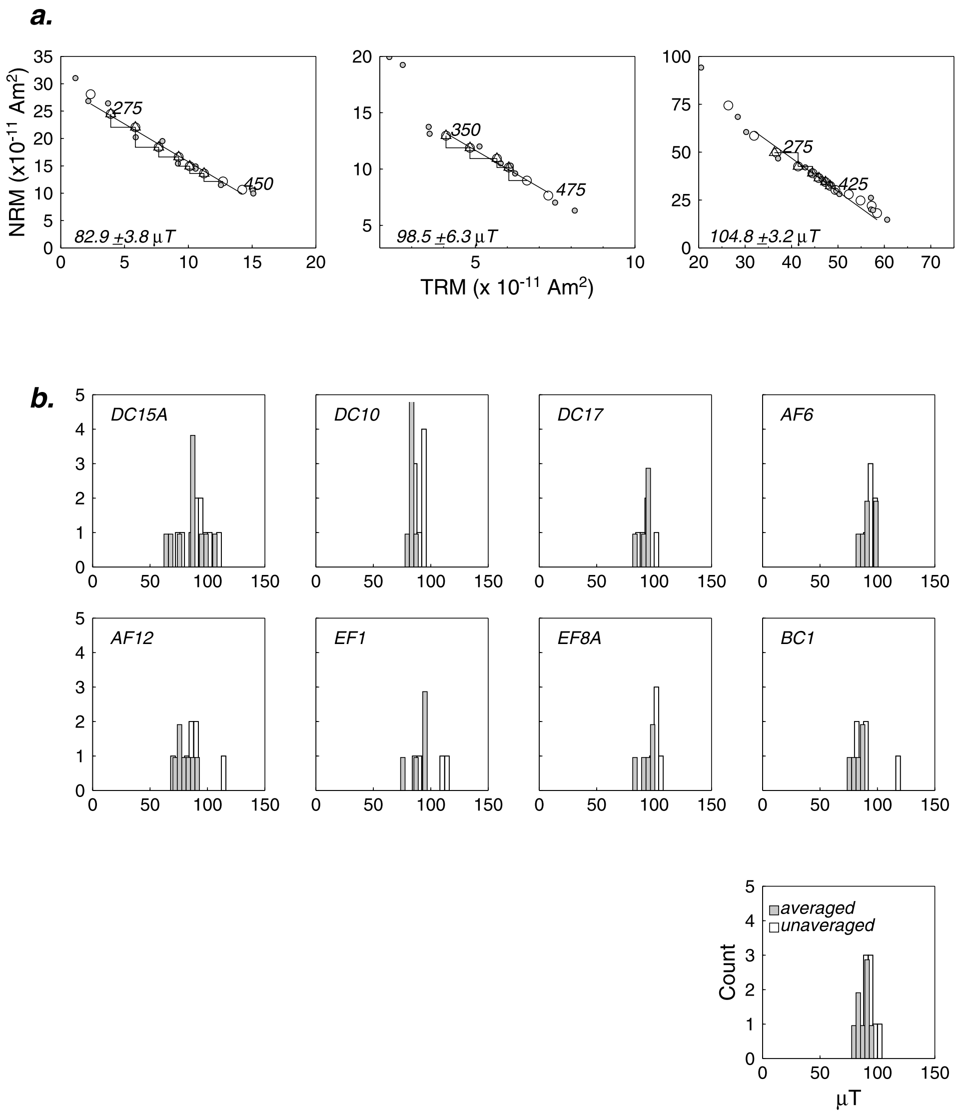

Figure 6. Paleointensity results from plagioclase crystals from the ~95 Ma Strand Fiord lavas (high Canadian Arctic) [after *Tarduno et al.*, 2002a]. (a) Typical plots of NRM versus TRM. The pTRM checks are shown by triangles. Large open circles are three-point sliding-window averages of raw data (small circles). Labeled points are temperatures (°C) representing the range used to determine paleointensity. (b) Summary of paleointensity data meeting reliability criteria from Arctic lavas. Abbreviations are AF, Agate Fiord; EF, Expedition Fiord; DC, Dragon Creek; and BC, Blacknose Ridge. Averaged data are shown as shaded; unaveraged data are shown as unshaded.

average field during the last 200 million years [e.g., *Juarez et al.*, 1998; *Selkin and Tauxe*, 2000].

[40] A similar study has been conducted on the Strand Fiord lavas of the high Canadian Arctic, which were also erupted during the Cretaceous Normal Polarity Superchron, at ~95 Ma [*Tarduno et al.*, 1997, 1998]. Fifty-one of the 110 crystals measured from eight lava units met the selection criteria. As in the case of the Rajmahal Traps these units are thought to average secular variation because geological indicators bear testament to time, and the angular dispersion of paleomagnetic directions recorded by whole rock samples matches that of a larger data set that spans the entire sequence [*Tarduno et al.*, 2002a]. Together these data yield a mean intensity of 93.8 ± 5.2 μT (1σ error) (Figure 6). This value is not significantly different from that obtained fitting the unaveraged data (93.9 ± 4.3 μT). The principal effect of averaging is to reduce within-flow scatter. The paleomagnetic dipole moment of $12.7 \pm 0.7 \times 10^{22}$ Am^2 suggested by the paleointensity data from the Arctic lavas agrees with the value from the Rajmahal Traps and indicates that high paleointensity of the latter rocks was not an isolated event within the Cretaceous Normal Polarity Superchron.

5.2. Submarine Basalt Glass Versus Single Plagioclase Crystals

[41] A high field intensity during the Cretaceous Normal Polarity Superchron has also been reported from continued analyses of submarine basaltic glass by *Tauxe and Staudigel* [2004]. This work supersedes prior studies using this material [e.g., *Pick and Tauxe*, 1993; *Selkin and Tauxe*, 2000], which

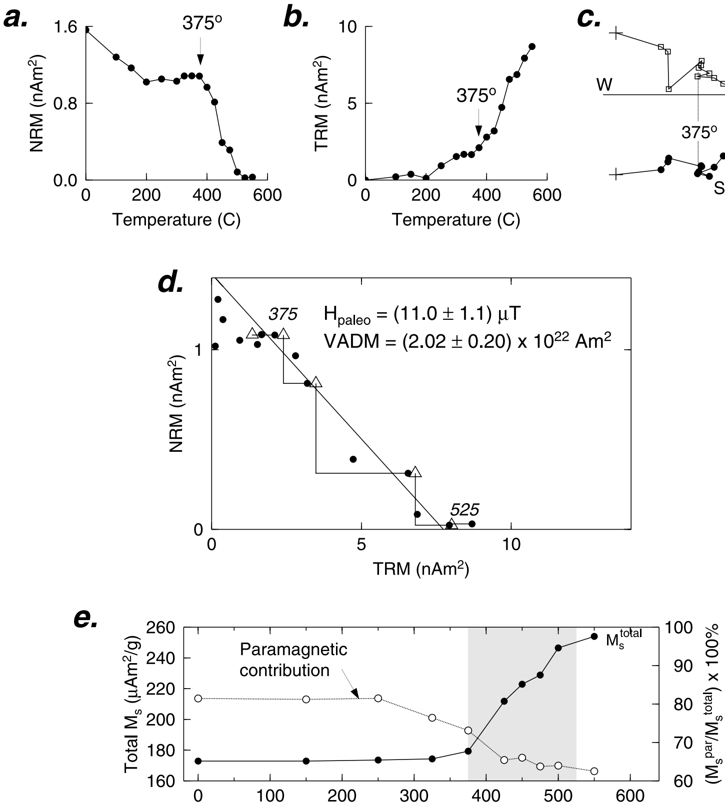

Figure 7. Example of Thellier paleointensity and rock magnetic experiments on submarine basaltic glass [from *Smirnov and Tarduno*, 2003]. Sample represented is from 621.06 m below seafloor (mbsf) at Ocean Drilling Program (ODP) Site 1203 [*Tarduno et al.*, 2002b]. (a) Thermal demagnetization of the natural remanent magnetization (NRM). (b) Partial thermoremanent magnetization (pTRM) acquisition. (c) Orthogonal vector plot of field-off steps for the unoriented sample (vertical projection of NRM, squares; horizontal projection of NRM, circles). (d) NRM-TRM plot (circles). Triangles are pTRM checks. (e) Magnetic hysteresis values as a function of Thellier heating temperature on a split of the submarine glass sample from 621.06 mbsf. The total saturation (M_s^{total}) before slope correction is shown (solid circles, solid line). Paramagnetic contribution is the difference (M_s^{par}) between M_s^{total} and the saturation magnetization after slope correction, normalized to M_s^{total} (open circles, dashed line). The systematic increases in M_s^{total} (shaded box) corresponds to the linear portion of the NRM/TRM plot that would normally be used to calculate paleointensity. These magnetic changes result in an artificially low paleointensity (values shown in Figure 7d). Figure 7 is reprinted from *Smirnov and Tarduno*, 2003], with permission from Elsevier.

concluded that the Superchron field was weak or of average strength. However, the new values reported from submarine basaltic glass are still some 30–40% lower, and more variable, than the field values based on plagioclase crystals.

[42] *Smirnov and Tarduno* [2003] observed neocrystallization of magnetite in Thellier heatings of submarine basaltic glass from three Cretaceous Ocean Drilling Program (ODP) sites (Figure 7). This alteration resulted in artificially low paleointensity values. *Smirnov and Tarduno* [2003] advocated the use of monitor specimens (i.e., subsamples) that would undergo the same Thellier heatings as the glass sample used for the paleointensity experiment. In this approach, magnetic hysteresis data are used to check for new magnetic mineral growth [see also *Haag et al.*, 1995]. *Smirnov and Tarduno* [2003] also demonstrated that low-temperature data, and the changing manifestation of the Verwey transition, could be used to detect changes (i.e., magnetite formation) in monitor glass specimens. *Riisager et al.* [2003a] used magnetic hysteresis experimental checks on monitor specimens and documented alteration during Thellier experiments on submarine basaltic glass from additional Cretaceous ODP sites.

[43] Glasses are thermodynamically unstable. With time and/or increased temperatures they become crystalline. The chemical bonds in glass break over a temperature transitional range [*Bouska*, 1993; *Donth*, 2001], and this range corresponds with temperatures applied during typical Thellier experiments. Therefore the tendency of older submarine basaltic glass to record lower and more variable paleointensity values could reflect subtle experimental alteration, enhanced by natural weakening of bonds, lowering of the glass transition temperature, and hydration with age.

5.3. Secular Variation and Field Morphology During the Cretaceous Normal Polarity Superchron

[44] The data from the Rajmahal Traps and Arctic basalts support a correlation between high field strength and low reversal frequency, as suggested in early models, particu-

larly the landmark work by *Cox* [1968]; see also discussion by *Banerjee* [2001]. Such a correlation is also supported by some considerations of core-mantle boundary processes [e.g., *Larson and Olson*, 1991]. However, what was the geometry of the field and the variation in directions? The high paleolatitude Arctic site, together with lower paleolatitude North American sites, forms a transect that is sufficiently long (>3700 km) that it allows a test of field geometry by examining paleomagnetic data from a single craton.

[45] This can be done by assuming that the paleomagnetic inclination observed at all sites records a dipole with progressively higher amounts of octupole contributions. The observed inclinations (I_o) with octupole (g_3^o or $G3$) terms are related to the expected colatitude (θ_e) of the sample sites by

$$\tan I_o = \frac{2\cos\theta_e + G3(10\cos^3\theta_e - 6\cos\theta_e)}{\sin\theta_e + G3\left(\frac{15}{2}cos^2\theta_e\sin\theta_e - \frac{3}{2}\sin\theta_e\right)}. \quad (3)$$

[46] After solving for θ_e and summarizing in pole space the grouping of the data provides information on which field morphology is most consistent with the observations. This test (Figure 8) indicates that the dipole alone provides the best fit, indicating that the time-averaged mid-Cretaceous field lacked significant octupolar or quadrupolar contributions [*Tarduno et al.*, 2002a].

[47] Another important way to examine the field is to examine the angular dispersion of the paleomagnetic data versus latitude [e.g., *Cox*, 1970]. Paleosecular variation (PSV) is typically defined by the angular dispersion of the lava mean directions, S:

$$S^2 = \frac{1}{N-1}\sum_{i=1}^{N}\Delta_i^2, \quad (4)$$

where N is the number of virtual geomagnetic poles (VGPs) and Δ_i is the angle between the ith VGP and the mean VGP. *McFadden et al.* [1991] updated and expanded upon early PSV analyses, proposing a model (model G) where the field is composed of two independent families: a dipole (or antisymmetric) family, which is linear with respect to latitude ($S_p = b\lambda$), and a quadrupole (or symmetric) family that is constant with latitude ($S_s = a$). The total VGP angular dispersion (S_λ) is given by

$$S_\lambda = \sqrt{(S_s)^2 + (S_p)^2}. \quad (5)$$

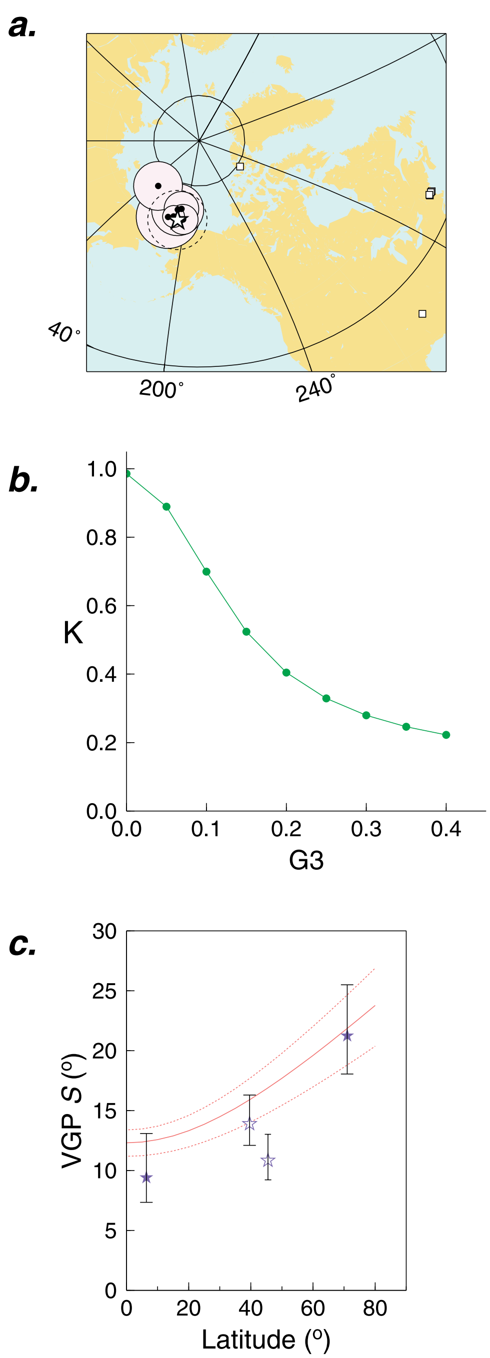

Figure 8. Tests of the morphology and secular variation of the geomagnetic field during the Cretaceous Normal Polarity Superchron [after *Tarduno et al.*, 2002a]. (a) Paleomagnetic pole (star) calculated from this direction data shown versus mid-Cretaceous poles of the North American craton (reference poles A–F (squares) from *Tarduno and Smirnov* [2001]). (b) Normalized estimate of precision parameter (K) for mid-Cretaceous North American poles and the high Canadian Arctic (assuming a 12° vertical axis rotation of the Arctic site, see discussion of *Tarduno et al.* [2002a]). Values shown assume various contributions of an octupole relative to the dipole field (G3). The poles are best grouped when no octupole (or quadrupole) component is included, highlighting the dominance of the axial dipole in the mid-Cretaceous time-averaged geomagnetic field. (c) Angular dispersion (S) of time-averaged virtual geomagnetic poles (VGP) and select data sets from the Cretaceous Normal Polarity Superchron (Northern Hemisphere (solid stars) and Southern Hemisphere (open stars)) shown with 95% confidence intervals [*Cox*, 1969]. Also shown is fit of secular variation model G [*McFadden et al.*, 1991] to 0–5 Ma data [*Johnson and Constable*, 1996] (solid curve) and 95% confidence interval (dashed curves).

[48] Although complete independence of the two families is unlikely [*Johnson and Constable*, 1996], the model provides an approximation useful for categorizing the past field. *McFadden et al.* [1991] proposed that the families varied systematically with respect to reversal rate during the last 160 million years. In particular, the smallest contribution of the quadrupole family was noted for the lowest rate of reversals, which was represented by an interval centered on the Cretaceous Normal Polarity Superchron.

[49] This model has been reassessed, using new data and a more stringent data selection of older data [*Tarduno et al.*, 2002a]. Databases were searched for paleomagnetic studies of lavas with ages between 83 and 120 Ma (conservative limits of the Cretaceous Normal Polarity Superchron) and having demagnetization treatment to select directions. This search yielded over 70 published studies. Next the following selection criteria were applied: (1) Radiometric ages, stratigraphic evidence, and magnetic polarity must be compatible with formation during the Cretaceous Normal Polarity Superchron. (2) Results must be based on thermal and alternating field demagnetization with principal component analyses used to fit directions. (3) More than five flows must be available with multiple samples ($N > 2$) for each flow. (4) The results should be free of evidence for remagnetization. (5) Paleolatitudes should not be controversial, and the tectonic setting should be sufficiently simple such that local faulting is not a potential source of intersite scatter. Furthermore, VGPs with latitudes less than 55° (a common cutoff for excursional behavior) were excluded.

[50] Paleomagnetic results meeting these requirements have been reported from only the Rajmahal Traps [*Klootwijk*, 1971; *Tarduno et al.*, 2001], the flood basalts of Madagascar [*Riisager et al.*, 2001], the Mount Carmel basalts of Israel [*Ron et al.*, 1990], and the high Arctic [*Tarduno et al.*, 2002a]. Fortunately, these studies represent high-, intermediate-, and low-latitude sites that can trace a basic description of VGP dispersion. These values can be compared with an analysis of paleomagnetic data from 0–5 Ma lavas [*Johnson and Constable*, 1996]. Results from the Northern and Southern hemispheres in the 0–5 Ma data are averaged because it is as yet unclear whether the hemisphere asymmetry observed reflects the field or differential data quality; a large data collection effort is currently in progress which will improve our resolution of the 0–5 Ma field [e.g., *Tauxe et al.*, 2004].

[51] The S values for the Cretaceous Normal Polarity Superchron data considered here at low latitudes and midlatitudes are less than those of the 0–5 Ma data (Figure 8). These comparisons support the inference that the contributions of the quadrupole family to the field were smaller during the Cretaceous Normal Polarity Superchron than during the field of the last 5 million years [*McFadden et al.*, 1991]. Secular variation studies are not sufficiently abundant to allow a continuous analysis for times older than 160 Ma. However, *Rochette et al.* [1997] discuss data supporting a lower quadrupole family contribution for the other well-defined superchron of the reversal chronology, the Kiaman Reversed Polarity Superchron (~262 to 320 Ma) [*Irving and Parry*, 1963; *Opdyke and Channell*, 1996].

5.4. Dipole Strength and Variation During a Time of Moderate Reversal Frequency

[52] The data sets discussed above suggest the time-averaged field is strong and overwhelmingly dipolar when it is in a nonreversing superchron state. These apparent characteristics of the geodynamo can be probed further through analyses of plagioclase crystals from lavas formed during intervals when the field was reversing (mixed polarity intervals). Basalt cores available from ocean drilling at a few key sites can be used for such an investigation. In particular, 25 subaerially erupted lavas have been recovered at ODP Site 1205 on Nintoku Seamount in the northwestern Pacific Ocean [*Tarduno et al.*, 2002b]. Nintoku Seamount is part of the Hawaiian-Emperor trend; the persistent volcanism at the Hawaiian hot spot has resulted in the eruption of lavas of relatively consistent composition during the construction of this volcanic feature. The lavas recovered at Nintoku Seamount are ~56 million years old [*Tarduno et al.*, 2003; *Duncan and Keller*, 2004] and thus formed after the Cretaceous Normal Polarity Superchron when the field was in a reversing state.

[53] Continued rock magnetic investigations afford additional insight into the magnetic signatures of the plagioclase crystals versus bulk samples of these lavas. In particular, the first-order reversal (FORC) technique [*Pike et al.*, 1999; *Roberts et al.*, 2000] can be used to learn more about grain size distributions. In FORC diagrams the horizontal axis (H_c) is a gauge of microcoercivity, while the vertical axis (H_b) is a measure of the interaction field between magnetic grains. Neither the whole rocks nor the single plagioclase crystals from Site 1205 show large magnetic interactions (Figure 9). Some plagioclase crystals have FORC distributions with well-defined maxima; others show a somewhat larger microcoercivity range. These distributions differ dramatically from those of the whole rocks and are more favorable for paleointensity studies. The whole rocks show evidence for very low microcoercivities consistent with multidomain grains. Interestingly, a secondary peak in the FORC distribution for the whole rocks of Site 1205 may be traced to magnetic inclusions in small plagioclase phenocrysts within the bulk samples (Figure 9).

[54] NRM/TRM data from 44 of 86 plagioclase crystals measured from 11 lava flow units from Nintoku Seamount Site 1205 yield a mean field value of 42.8 ± 13.6 μT (1σ error) [*Tarduno and Cottrell*, 2005] (Figure 10). The plagioclase crystals come from lavas that span the Site 1205 basement sequence. We can be confident that time elapsed between the lavas, because some lava tops are deeply weathered. In other cases, soil horizons separate the lavas. Paleomagnetic directions, and the associated estimate of polar angular dispersion [*Tarduno et al.*, 2003], further indicate that the flows are independent in time and average secular variation. Thus the virtual dipole moments associated with each lava mean together make up a paleomagnetic dipole moment. This moment, 8.9×10^{22} Am^2 [*Tarduno*

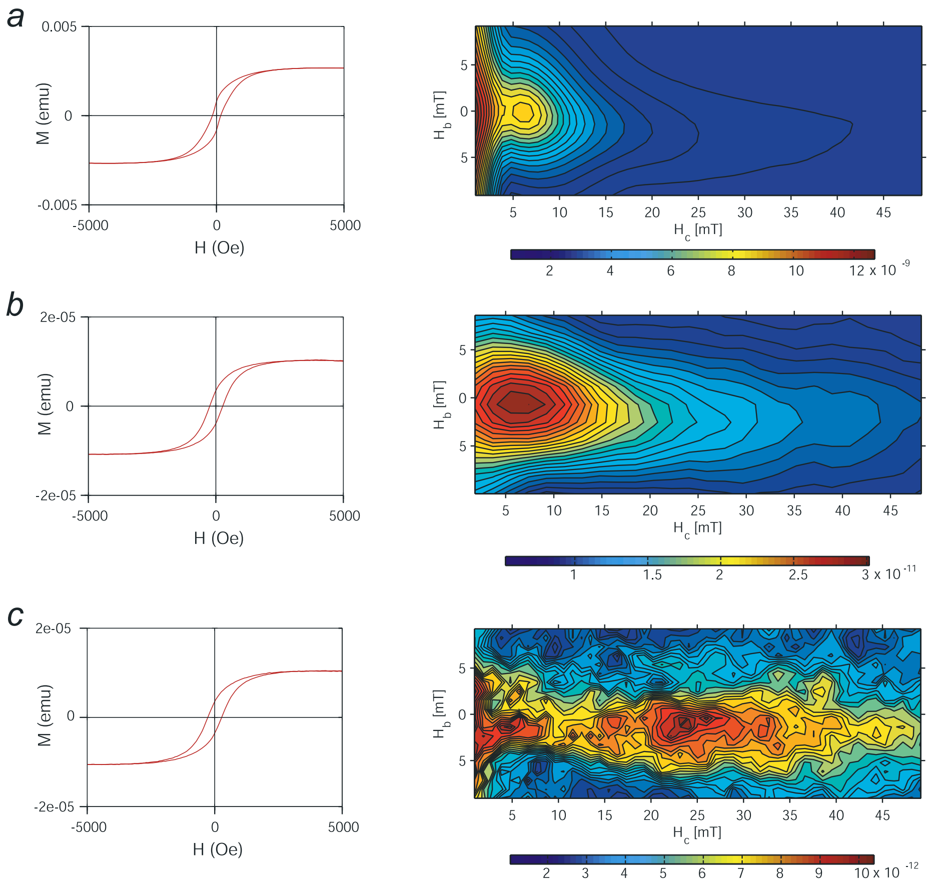

Figure 9. (a) (left) Magnetic hysteresis data for a whole rock basalt sample from ODP Site 1205 and (left) corresponding first-order reversal curve (FORC) [*Pike et al.*, 1999; *Roberts et al.*, 2000] diagram. (b and c) (left) Magnetic hysteresis data from single plagioclase crystals separated from Site 1205 lavas and (right) their corresponding FORC diagrams. All FORC diagrams use a smoothing factor [*Pike et al.*, 1999; *Roberts et al.*, 2000] of 5. Figures 9a and 9b are based on 50 curves; Figure 9c is based on 75 curves. Figure 9 is from *Tarduno and Cottrell* [2005].

and Cottrell, 2005], is significantly weaker (at the 95% confidence level) than those derived from Thellier analyses of plagioclase crystals from Cretaceous Normal Polarity Superchron lavas. The standard deviation of the Site 1205 virtual dipole moments (32%) is similar to that of the modern field but nearly 3 times greater than those recorded by the Cretaceous Normal Polarity Superchron plagioclase crystals.

5.5. Whole Rocks Versus Single Plagioclase Crystals Rematch: Role of Maghemitization

[55] A study of whole rock samples of the Site 1205 lavas by *Carvallo et al.* [2004a, 2004b] provides an opportunity to make another comparison between paleointensity data based on whole rock samples and single plagioclase crystals. *Carvallo et al.* [2004a] used rock magnetic tests to select lava samples that appeared to have only minor multidomain magnetic mineral components. However, further tests showed that samples from only one lava unit had escaped low-temperature oxidation. Plagioclases large enough for paleointensity measurements are not available from that aphyric basalt. However, the range of paleointensity values obtained from the whole rocks (34.2–39.4 μT) by *Carvallo et al.* [2004a] is compatible with that available from single plagioclase crystals.

[56] *Carvallo et al.* [2004a] also performed Thellier experiments on some of the Site 1205 lavas that had undergone low-temperature oxidation. Interestingly, these samples yielded nominal whole rock paleointensity values that are lower than those obtained using plagioclase crystals from the same units [*Tarduno and Cottrell*, 2005]. This observation is consistent with the hypothesis that low-temperature oxidation and the acquisition of CRM can lead to a low field bias in paleointensity data from whole rocks [*Tarduno and Smirnov*, 2004; *Smirnov and Tarduno*, 2005]. This may be an important causal factor for the low field values derived from whole rock lava samples older than 10 Ma (Figure 1).

5.6. Dipole Strength During an Interval of Very High Reversal Frequency

[57] The Late Jurassic appears to represent a time of very high reversal frequency [*Opdyke and Channell*, 1996], perhaps the highest of the last 200 million years [*Tivey et al.*, 2005]. Basalt recovered during drilling at ODP Site 801 [*Lancelot et al.*, 1990; *Plank et al.*, 2000] in the Pigafetta Basin of the western Pacific provides an opportunity to

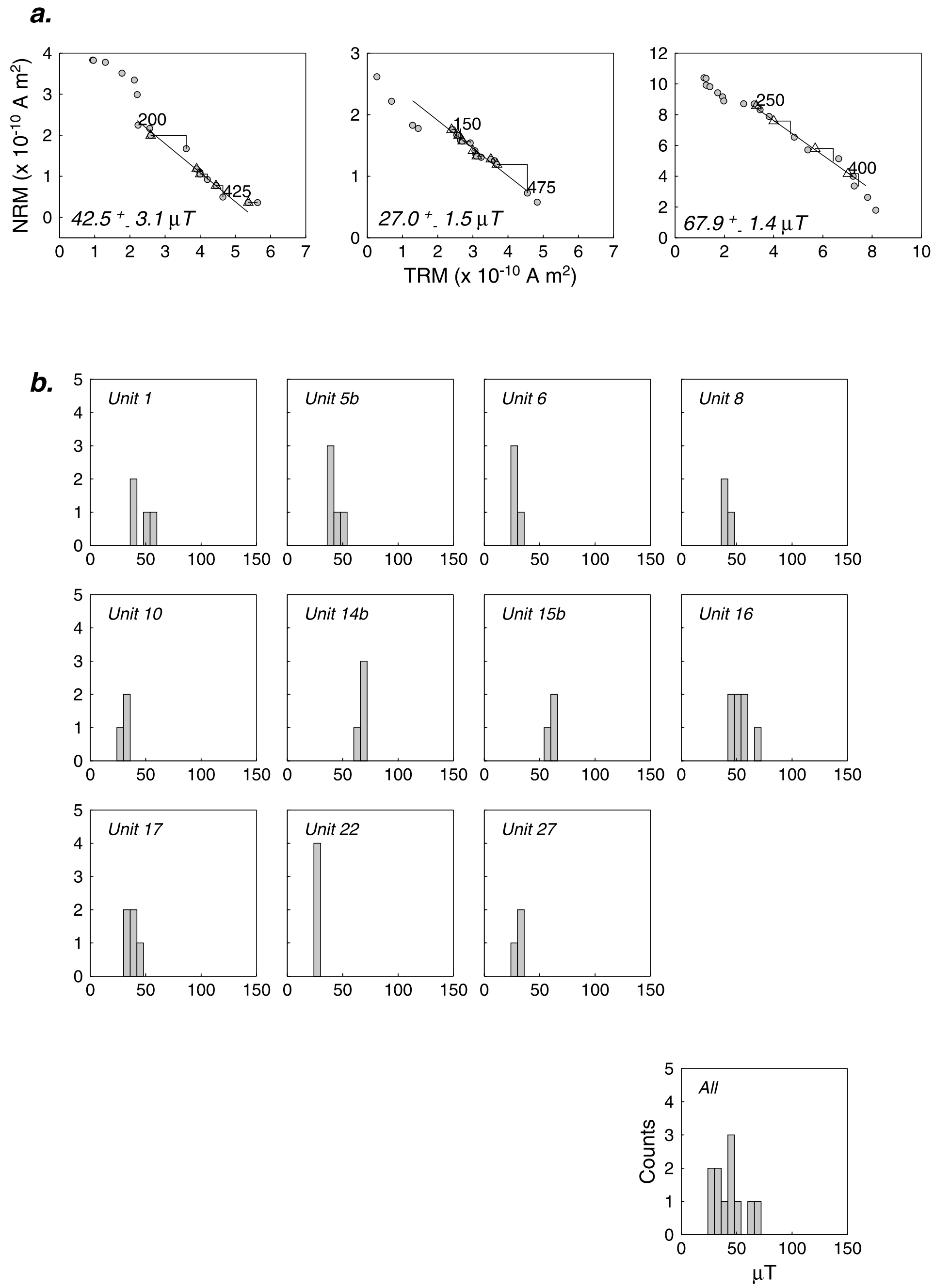

Figure 10. Paleointensity results from plagioclase crystals from the ~56 Ma Nintoku Seamount lavas (ODP Site 1205) [from *Tarduno and Cottrell*, 2005]. (a) Typical plots of NRM versus TRM (circles). The pTRM checks are shown by triangles. Labeled points are temperatures (°C) representing the range used to determine paleointensity. (b) Summary of paleointensity data meeting reliability criteria. Unit numbers are from *Tarduno et al.* [2002b, 2003].

examine this interval. The uppermost part of the basement sequence at this site consists of alkalic basalt with an age of ~160 Ma, constrained by $^{40}Ar/^{39}Ar$ geochronology [*Koppers et al.*, 2003].

[58] As in the case of the Site 1205 lavas we can apply the FORC technique to learn more about magnetic grain size distributions and interactions. Again, these data show clearly that single plagioclase crystals should be favored over bulk rock analyses. Plagioclase crystals from Site 801 basalts yield FORC data consistent with noninteracting single-domain-like carriers (Figure 11). FORC diagrams from Site 801 whole rock samples, however, suggest the presence of larger multidomain magnetic phenocrysts. Individual magnetic hysteresis curves from plagioclase are wasp-waisted (Figure 11), suggesting the presence of some superparamagnetic grains.

[59] Paleointensity values for the Site 801 plagioclase crystals have been isolated at temperatures >325°C, beyond the influence of viscous remanent magnetizations potentially held by ultrafine grains close to the single domain/super-

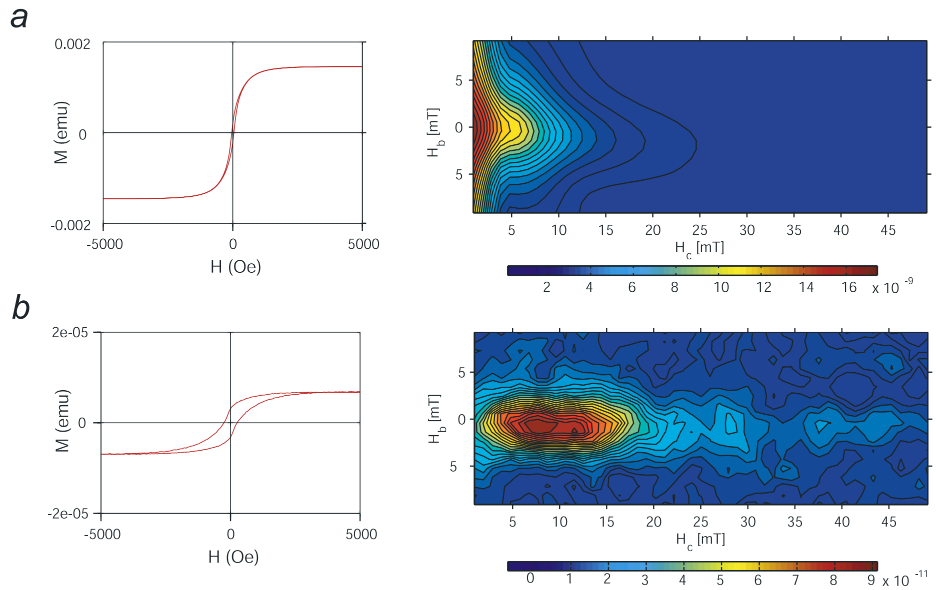

Figure 11. (a) (left) Magnetic hysteresis data for a whole rock basalt sample from ODP Site 801 and (right) corresponding FORC [*Pike et al.*, 1999; *Roberts et al.*, 2000] diagram. (b) (left) Magnetic hysteresis data from a single plagioclase crystal separated from an ODP Site 801 lava and (right) its corresponding FORC diagram. The FORC diagrams use a smoothing factor [*Pike et al.*, 1999; *Roberts et al.*, 2000] of 5. Fifty and 75 curves are used for the FORC diagrams of Figures 11a and 11b, respectively. Figure 11 is from *Tarduno and Cottrell* [2005].

paramagnetic boundary [*Dunlop and Özdemir*, 1997]. Twelve of 23 plagioclase crystals from three units of ODP Site 801 have yielded Thellier results meeting experimental selection criteria (Figure 12). Together, these values yield a mean of 21.0 ± 1.57 μT [*Tarduno and Cottrell*, 2005]. Calculation of virtual dipole moments (VDMs) is complicated by tectonic uncertainties at the site (dip of the lavas). Using paleolatitudes of 10° and 30°, which bound the available constraints [*Tivey et al.*, 2005; *Tarduno and Cottrell*, 2005], VDMs of 5.21 ± 0.39 × 10^{22} Am^2 and 4.11 ± 0.31 × 10^{22} Am^2, respectively, are suggested.

[60] These results are too limited to average secular variation. Nevertheless, the field values obtained are the lowest recorded to date by single plagioclase crystals. This is interesting in light of the field strength implied by the pattern of Pacific marine magnetic anomalies that correspond to the age of the alkalic basalt at Site 801. *McElhinny and Larson* [2003] have emphasized that the amplitude of these Late Jurassic marine magnetic anomalies suggests a weak geomagnetic field intensity.

5.7. Mean Field Strength Recorded by Basalt, Glass, and Feldspar Versus the Present Dipole Decay

[61] The studies described above of plagioclase feldspar provide a basic outline of geomagnetic field intensity during the last 160 million years. Before considering further the implications of these data, it is worthwhile to revisit one of our starting points: the low apparent field strength from older whole rock basalt samples (Figure 1). We now expand this discussion to include results from submarine basaltic glass.

[62] We examine these data with respect to two levels: 8 × 10^{22} Am^2, the strength of the present-day field, and 4 × 10^{22} Am^2 (Figure 13). The latter number is important because it is the strength level below which excursional field behavior is often observed from lavas erupted during the last 10 million years [*Tanaka et al.*, 1995]. A reduction of the present dipole to this value would result in the emergence of large nondipole field foci at many locations [*Guyodo and Valet*, 1999].

[63] For the whole rock lavas we select only those studies which report Thellier results from multiple lava flows (≥3) [*Riisager et al.*, 2002]. Hence this selection differs from the data of Figure 1, where all lava results were included. In many studies, information is unavailable to determine whether the data adequately sample secular variation such that the average can be considered a paleomagnetic dipole moment (the minimal requirement on the number of lava flows connotes that some values are not). In other studies, there are demonstrable problems associated with low-temperature oxidation/maghemitization (e.g., data of *Tanaka and Kono* [2002], used by *Zhao et al.* [2004]). For the 133 Ma interval (the Paraná basalt), two results are shown. In the study yielding the lower paleointensity result [*Kosterov et al.*, 1998], chemical and/or physical changes during the Thellier experiments were noted.

[64] Regardless of the problems, the whole rock data shown in Figure 3 are those that are typically used in evaluating the past field [e.g., *Goguitchaichvili et al.*, 2002]. They do indeed show several striking characteristics: There is no clear pattern of paleointensity with respect to geomagnetic reversal rate, and the mean field value is rather low. During the Cretaceous Normal Polarity Superchron, when directional data from lavas indicate an overwhelmingly dipolar field [*Tarduno et al.*, 2002a], the field strength

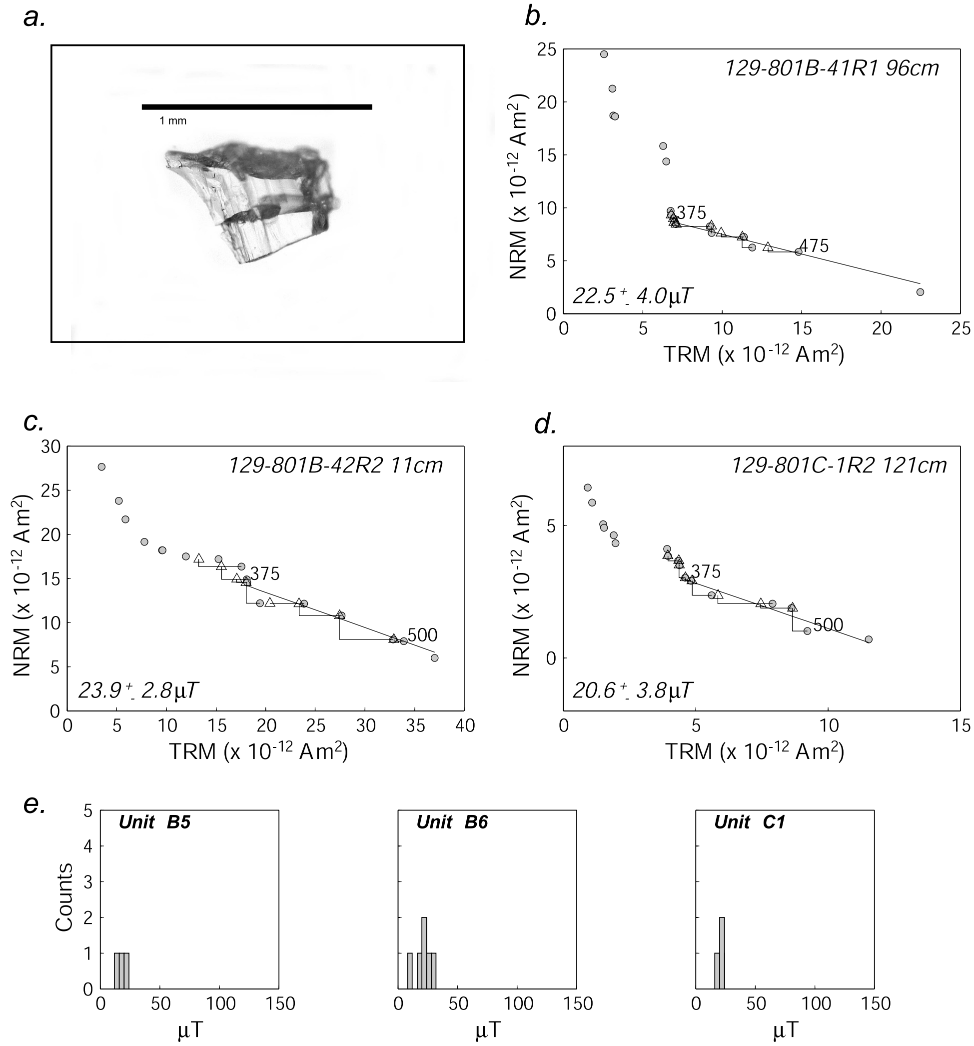

Figure 12. (a) Typical plagioclase crystal from ODP Site 801 (Pigafetta Basin) used for Thellier paleointensity experiments. (b) Thellier data for plagioclase crystal shown in Figure 12a. NRM versus TRM (circles) with pTRM checks (triangles) are shown. Labeled points are temperatures (°C) representing the range used to determine paleointensity. (c and d) Additional examples of Thellier data from other plagioclase crystals. (e) Summary paleointensity data. Figure 12 is from *Tarduno and Cottrell* [2005].

derived from studies of whole rock lavas is below 4×10^{22} Am^2.

[65] Many pillow lavas with associated glass can be erupted in a very short time period. In this case, individual results from submarine glass would sample, and overrepresent, the same instantaneous geomagnetic field value. Therefore it is important to assess whether the data are independent (as has been done for the other paleointensity data types). Unfortunately, time-averaged results are not available from most of the submarine basaltic database; only results from Thellier experiments on individual samples have been reported [*Selkin and Tauxe*, 2000]. Nevertheless, these are of very high technical quality, and some high paleointensity values are seen in the individual sample results for the Cretaceous Normal Polarity Superchron [*Tauxe and Staudigel*, 2004]. Yet the aggregate of individual sample results suggests an overall mean field strength for the last 160 million years that is even lower than that implied by the whole rock basalt data: It is close to half the present-day field value. In contrast, only during the ultrahigh reversal frequency interval of the Jurassic does the field strength recorded by plagioclase feldspar approach the 4×10^{22} Am^2 level.

[66] These radically different views on the ''mean value'' relate to the potential significance of modern field strength trends. In particular, dipole strength has decreased by over 10% during the last 150 years; if this trend were to continue, the dipole would fall to zero during the next 2000 years [e.g., *Leaton and Malin*, 1967; *McDonald and Gunst*, 1968; *Heirtzler*, 2002]. This rapid rate of decay, combined with spatial patterns of the field available from satellite data, has fueled speculation that we are in the early phases of a geomagnetic field reversal [*Hulot et al.*, 2002; *Olson*, 2002]. If the mean field values of whole rock basalt or submarine basaltic glass are correct, the recent dipole decay could represent a trend toward the long-term mean from a

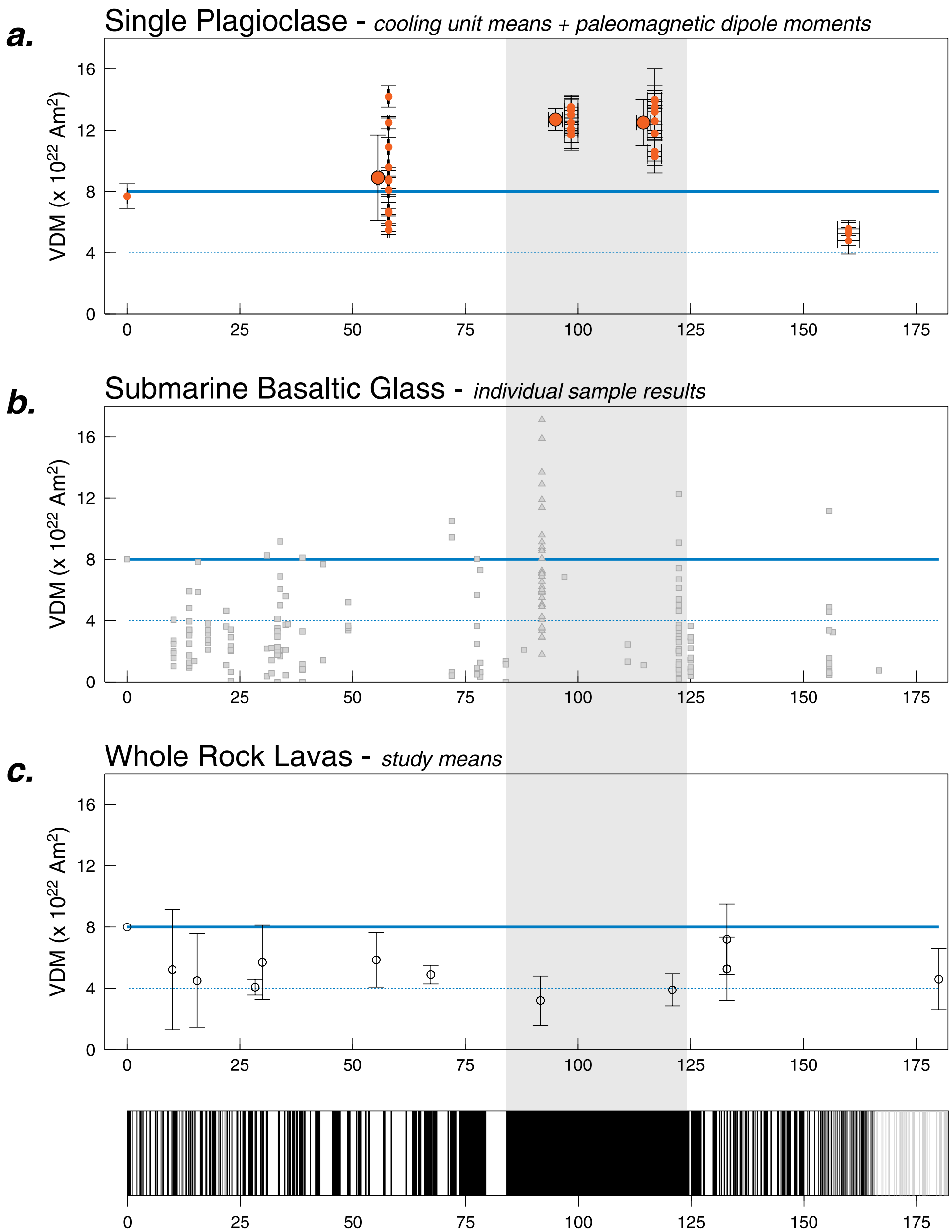

Figure 13. Paleointensity history for times older than 10 Ma based on Thellier analyses using different natural recorders. (a) Single plagioclase crystals [*Tarduno and Cottrell*, 2005]. Data represent cooling unit means (small circles) and paleomagnetic dipole moments (large circles). (b) Submarine basaltic glass [*Tauxe and Staudigel*, 2004]. Data are individual sample results (see text). (c) Whole rock basalt samples. Data plotted represent study means following criteria of *Riisager et al.* [2002]: at least nine determinations from three cooling units. Data sets plotted include 15–16 Ma [*Prévot et al.*, 1985], ~28 Ma (K-Ar age) [*Goguitchaichvili et al.*, 2001], 30 Ma [*Riisager et al.*, 1999], 55.2 Ma [*Riisager et al.*, 2002], ~67 Ma (K-Ar age) [*Goguitchaichvili et al.*, 2004], ~92 Ma (K-Ar age) [*Zhao et al.*, 2004], ~121 Ma (K-Ar age) [*Rixiang et al.*, 2001], 133 Ma [*Goguitchaichvili et al.*, 2002], 133 Ma [*Kosterov et al.*, 1998], and 180 Ma [*Kosterov et al.*, 1997].

short-term, unusually high field value; therefore the recent decay would carry no special significance. In fact, on the basis of results from submarine basaltic glass, *Tauxe and Staudigel* [2004] conclude that we are in, or heading toward, a superchron state.

[67] We hold an antithetical view. The paleointensity results from plagioclase crystals imply a higher mean field strength, closer to the present-day field value and that recorded by the youngest samples of all three data types. This higher mean implies that the present decay represents something more than a return to an average value. Thus the present-day rapid dipole decay could indeed herald a trend toward a reversal or excursion.

[68] Although the three data types suggest different mean field values and histories, the comparative rock magnetic and paleointensity experiments discussed previously pro-

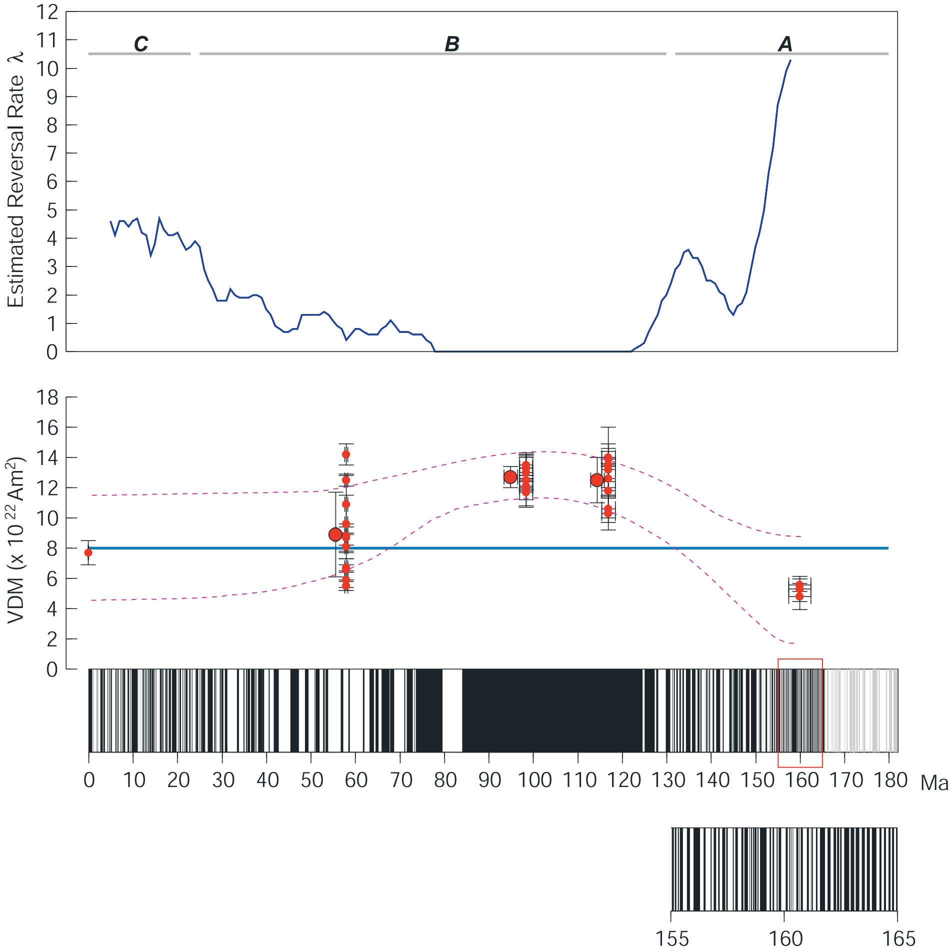

Figure 14. Geomagnetic reversal timescale [*Opdyke and Channell*, 1996; *Gradstein et al.*, 2004], Thellier paleointensity results from analyses of single plagioclase crystals and estimates of reversal rate for the last 180 million years. Figure 14 is from *Tarduno and Cottrell* [2005]. Inset shows an expanded version of the 165–155 Ma reversal chronology. Small circles are virtual dipole moments (VDM) (cooling unit means of multiple single crystal results), whereas large symbols are averages of the VDMs, representing paleomagnetic dipole moments, with their 1σ uncertainty regions. Paleointensity data are from *Cottrell and Tarduno* [1999, 2000], *Tarduno et al.* [2001, 2002a], and *Tarduno and Cottrell* [2005]. Dashed line shows variation of paleomagnetic dipole moment standard deviations consistent with the available data. Reversal rate curve is based on a 10 Myr sliding window. Also shown is the division of the reversal chronology into stationary and nonstationary intervals (A, B, and C) following *Gallet and Hulot* [1997].

vide a means for judging their relative accuracy. Rock magnetic investigations of whole rock lavas and plagioclase crystals indicate the former are more susceptible to natural and experimental alteration. Direct comparisons of paleointensity data indicate that this alteration leads to low field bias in the whole rock data. Experiments using older submarine basaltic glass detail magnetite formation and low field bias. On the basis of these experiments we believe that the mean long-term field strengths derived from whole rock basalts and submarine glass are incorrect but that in some cases relative field strength trends may be preserved within a data type (e.g., the older submarine glass results). Below we proceed with interpretations of the paleointensity data from plagioclase crystals, data that we feel best represent ancient field strengths because of their resistance to natural and experimental change.

5.8. Toward a Holistic View of the Field and Its Relationship to Mantle Convection

[69] Although significant temporal variations of field behavior during the Cretaceous Normal Polarity Superchron cannot be excluded (particularly near its edges), a stable dipole-dominated field is most consistent with available time-averaged intensity data from plagioclase data and global directional data (Figure 14). The further analyses of directions from whole rock samples of lavas from the same stratigraphic sequences from which the paleointensity data were derived suggest that geomagnetic reversals, field morphology, secular variation, and intensity, which are sometimes considered to be isolated phenomena, may instead be related on timescales of thousands [e.g., *Love*, 2000] to hundreds of millions of years [*Cox*, 1968; *Irving and Pullaiah*, 1976].

[70] The data available from plagioclase crystals from lavas erupted prior to and after the Cretaceous Normal Polarity Superchron provide further context for these findings. The Site 1205 lavas were erupted during a time of moderate reversal frequency (<1 reversal Myr^{-1}). The Site 801 alkalic lavas were formed when the occurrence of reversals was very high (>10 reversals Myr^{-1}). Both assessments employ the updated timescale of *Gradstein et al.* [2004]. When combined, the plagioclase-based paleointensity data assay the range of reversal behavior displayed by the geodynamo during the last 200 million years. The pattern suggested by the ensemble of single crystal paleointensities traces an inverse relationship between field strength and reversal frequency.

[71] There are currently several competing interpretations of the geomagnetic reversal chronology that provide a geophysical backdrop for further considerations of the paleointensity data based on plagioclase crystals. One view is that nonstationary intervals bound the Cretaceous Normal Polarity Superchron [*McFadden et al.*, 1991]. Another view is that the history is best represented by a series of stationary regimes [*Lowrie and Kent*, 2004]. Still another interpretation suggests that the Cretaceous Normal Polarity Superchron is part of a nonstationary interval spanning 130–25 Ma, followed by a stationary interval [*Gallet and Hulot*, 1997]. The latter interpretation has led to the suggestion that superchrons may reflect nonlinear dynamo processes [*Hulot and Gallet*, 2003].

[72] In contrast, superchrons may reflect times when the nature of core-mantle boundary heat flux allows the geodynamo to operate at peak efficiency, as suggested in some numerical models [*Glatzmaier et al.*, 1999; *Roberts and Glatzmaier*, 2000], whereas the succeeding period of reversals may signal a less efficient dynamo with a lower and more variable dipole intensity. This interpretation is qualitatively most similar to the observations based on Thellier analyses of single plagioclase crystals available to date. The timescale for the transition between these states is consistent with a lower mantle control on the geodynamo. It is difficult to see how this could be accomplished through the sluggish, isolated lower mantle of some views of Earth structure [*Anderson*, 2004], and it is instead more compatible with the deep penetration of slabs that can eventually influence core-mantle boundary processes. It is also important to note that the Cretaceous Normal Polarity Superchron correlates with a host of phenomena that indicate an unusually vigorous mantle, including motion between groups of hot spots [*Tarduno and Gee*, 1995; *Tarduno and Smirnov*, 2001; *Antretter et al.*, 2002; *Riisager et al.*, 2003b], and the emplacement of Ontong Java [*Tarduno et al.*, 1991; *Mahoney et al.*, 2002] and other large oceanic plateaus [*Larson*, 1991; *Coffin and Eldhom*, 1994].

6. PRECAMBRIAN AND INNER CORE GROWTH

[73] There is considerable disagreement in models for the thermal evolution of the Earth. Most generally predict that the solid inner core started to grow no earlier than approximately 2.5–2.7 billion years ago [e.g., *Labrosse et al.*, 2001; *Labrosse*, 2003; *Labrosse and Macouin*, 2003; *Buffett*, 2003], but some prefer a greater age [e.g., *Gubbins et al.*, 2004]. In this regard, older Archean rocks are of prime interest for paleomagnetic and paleointensity investigations because they may record a time when the familiar compositionally driven convection of the modern dynamo was not operating and a solid inner core [*Hollerbach and Jones*, 1993] did not play a role in controlling the geometry of outer core flow.

[74] Unfortunately, as we move to rocks of the Precambrian, the number of pristine lava flow sequences available for study drops precipitously. Although there are a few exceptions, we must turn instead to dikes and other intrusive bodies for the Proterozoic to Archean interval [*Dunlop and Yu*, 2004; *Macouin et al.*, 2004]. For mid-Archean and older times, plutonic rocks are most common; the mafic rocks available (“greenstones”) often have seen the massive formation of secondary magnetite associated with serpentinization. Below we review the challenges inherent to the study of both time intervals, progress made in using the magnetization of single silicate grains, and potential for further advances.

6.1. Proterozoic to Archean

[75] Mafic dikes and sills of Proterozoic to Archean age are exposed on several continents, and these have been targeted in many previous paleomagnetic investigations [e.g., *McElhinny and Opdyke*, 1964; *Fahrig et al.*, 1965; *Irving and Yole*, 1972; *Buchan and Halls*, 1990]. These contain feldspars that could carry magnetic inclusions better suited for paleointensity analysis than bulk rock samples, as is the case for the younger lavas discussed in section 4. *Smirnov et al.* [2003] explored this possibility in a study of mafic border dikes associated with the 2.45 Ga Burakovka intrusion of the Karelian craton (Russia). Thin border dikes were the focus of study so that the effects of cooling rate were minimized. Clear feldspars were selected for study from the dikes; in other Proterozoic-Archean dike provinces, clouded feldspars reflecting exsolution are common [*Halls and Zhang*, 1998]. Magnetic hysteresis parameters indicate multidomain-like behavior of whole rock samples and pseudosingle- to single-domain behavior for plagioclase crystals separated from the Karelian dikes. Whole rock samples also show a small anisotropy as recorded by systematic variations in magnetic hysteresis data, presumably recording a flow fabric within the dikes. No significant anisotropy was observed in the plagioclase crystals. As in the study of younger rocks discussed in section 4, the lack of anisotropy is important because it indicates that there is not a dominant preferred alignment of elongated particles in the feldspars that could bias TRM acquisition in Thellier experiments.

[76] Transmission electron microscopy analyses suggest that the magnetic inclusions in the plagioclase are equant to slightly elongated and range in size between 50 and 250 nm. Thermal demagnetization data of a saturation remanence imparted on single plagioclase crystals at 10 K are charac-

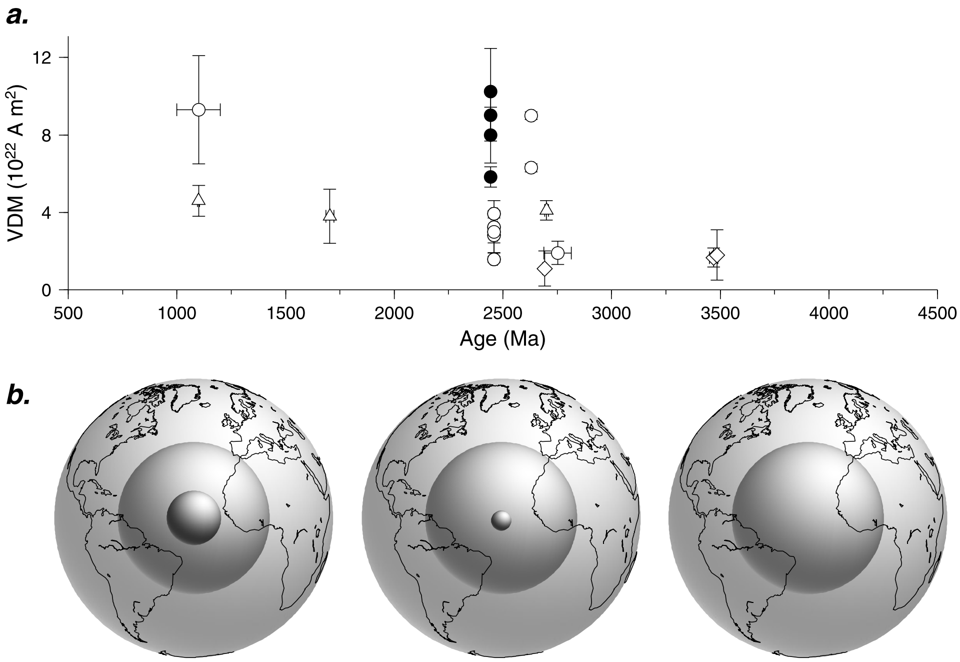

Figure 15. (a) VDM data for the Precambrian based on Thellier analyses (1σ uncertainties shown). Triangles are results from slowly cooled whole rocks; results from whole rock dikes are shown by open circles. Data shown are those selected by *Smirnov et al.* [2003], supplemented with results from the the Stillwater Complex [*Selkin et al.*, 2000] and the Matachewan dikes [*Macouin et al.*, 2003]. The Matachewan dike whole rock values obtained from standard Thellier analyses agree with those derived using microwaves [*Halls et al.*, 2004]. Diamonds are from whole rock samples of komatiites [*Hale*, 1987; *Yoshihara and Hamano*, 2004]; the secondary magnetic minerals they carry do not record true thermoremanent magnetizations. Solid circles are results from feldspars separated from border dikes of the Burakovka intrusion. The paleointensity derived from these crystals, as well as secular variation data derived from whole rocks [*Smirnov and Tarduno*, 2004], suggests that geomagnetic field behavior in Late Archean-Early Proterozoic was similar to that of the last 5 million years. This further suggests that inner core growth commenced by 2.5 Ga. (b) Hypothetical history of inner core growth. Figure 15b is after *Smirnov et al.* [2003].

terized by a well-defined Verwey transition at ~120 K indicating the presence of stoichiometric magnetite, similar to that reported from whole rock samples.

[77] These rock magnetic data demonstrate the feasibility of the single plagioclase paleointensity approach as applied to select rocks of the Karelia and Proterozoic-Archean rocks of other cratons. Results from only four dikes are available to date from Karelia, and it cannot be expected that these adequately record the full range of geomagnetic secular variation. Nevertheless, the Thellier paleointensity data are within the range of modern field values (Figure 15). Secular variation and field morphology during this interval also look amazingly similar to that of the last 5 million years [*Smirnov and Tarduno*, 2004] (Figure 16). In fact, select mean values from sites in the Superior Province [*Smirnov and Tarduno*, 2004], the Karelia Craton [*Mertanen et al.*, 1999], and Pilbara Craton [*Strik et al.*, 2003] suggest a somewhat lower low-latitude angular dispersion than that of the 0–5 Ma field, hinting at a more dipole-dominated field in Late Archean-Early Proterozoic times. The data are limited in number, and such an inference is speculation. Nevertheless, it is interesting that some numerical dynamo models incorporating a smaller inner core [e.g., *Roberts and Glatzmaier*, 2001] predict a morphology that is more dipolar than the modern field.

6.2. Multidomain Magnetic Mineral Carriers in Proterozoic-Archean Dikes

[78] A recent signal study of mineral fractions from Matachewan dikes of the Superior Province (Canada) by *Dunlop et al.* [2005] underscores the expected differences in paleointensity recording fidelity of whole rocks and feldspars. An extract of mafic minerals (pyroxene grains, with minor amounts of amphibole, biotite, and magnetite) displayed multidomain characteristics and nonlinear behavior during Thellier experiments. Concave NRM/TRM plots, reflecting the self-demagnetization response of the multidomain grains, were observed. Furthermore, pTRM checks were ineffectual at distinguishing alteration because the magnetization of multidomain carriers is sensitive to magnetic history. As outlined by *Dunlop et al.* [2005], this leads to an interesting experimental dilemma.

[79] The failure of a pTRM at very high temperature (e.g., >550°C) during a Thellier experiment using bulk samples from these dikes might lead one to conclude that alteration has occurred and that paleointensity estimation should should be limited to data obtained at lower temper-

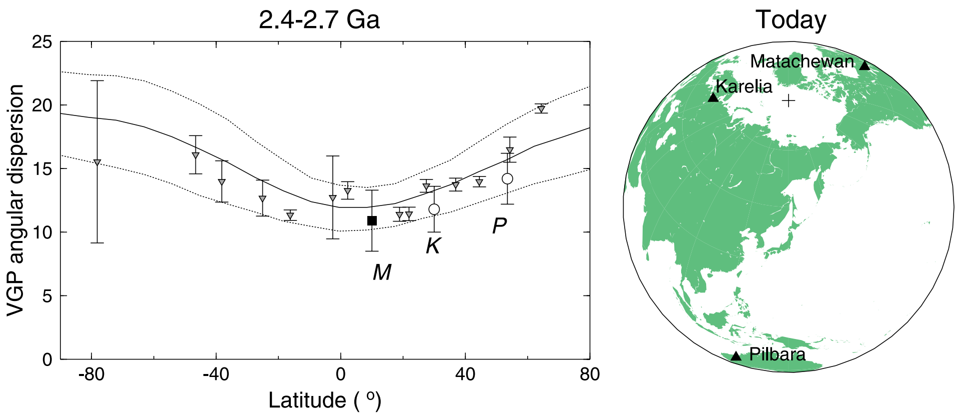

Figure 16. Archean-Proterozoic secular variation after *Smirnov and Tarduno* [2004]. Latitudinal dependence of angular dispersion of virtual geomagnetic poles (VGPs) based on paleomagnetic data from the Karelia craton (K) [*Mertanen et al.*, 1999], the Pilbara craton (P) [*Strik et al.*, 2003], and the Superior Province (Matachewan dikes (M)) [*Smirnov and Tarduno*, 2004]. Confidence limits on K, P, and M are 1σ and were calculated using a N-1 jackknife [*Efron*, 1982]. These estimates are shown with VGP angular dispersion estimates for lavas of the last 5 million years [*Johnson and Constable*, 1996] and a model modified from *Constable and Parker* [1988] (solid line) and its 95% confidence interval [from *Johnson and Constable*, 1996]. Also shown is the present-day location of sites.

atures. An investigator might be persuaded by the pTRM failure to select a relatively low-temperature portion of the NRM/TRM curve for paleointensity calculation, safely avoiding the effects of experimental alteration. Or only data obtained at high temperatures (but just below the temperature where pTRM-checks failed) might be chosen, because it is these data that would be expected to be least contaminated by secondary magnetizations. However, the pTRM failure might simply reflect changes in the initial state of multidomain magnetic carriers, carriers which fail to meet the demands of the Thellier paleointensity method. In this case the first choice (lower-temperature data) would result in an overestimate of past field strength, whereas the second choice (high-temperate data) would result in a paleofield underestimate.

[80] In practice, gross nonlinearity, such as the concavity that leads to the paleointensity miscalculations described above, should be detected by visual inspection of the NRM/TRM plots and/or by additional tests (including pTRM tail checks [*Shcherbakova et al.*, 2000]). *Dunlop et al.* [2005] estimate that it would take no more than 20% contamination by multidomain grains before the response of single-domain carriers would be eclipsed. However, the relevant issue for our consideration is whether small multidomain effects are present, such that they cause bias in the data. All mafic dikes will contain the mafic components studied by *Dunlop et al.* [2005]. Therefore bias is a strong possibility. For example, some paleointensity results from whole rock samples of the Matachewan dikes are restricted to high temperatures and yield seemingly low field values [*Macouin et al.*, 2003].

[81] The alternative is to avoid the experimental dilemma described above altogether and select a better magnetic recorder than the whole rocks. *Dunlop et al.* [2005] also studied plagioclase feldspars separated from Matachewan dikes. The plagioclase selected for study were clouded, and the magnetic inclusions clearly reflect exsolution. Notwithstanding the issues that must be addressed when crystals with exsolved magnetic inclusions are used in paleointensity experiments (issues which are discussed in section 6.3 in our consideration of even older rocks), the rock magnetic tests showed a dramatic contrast with those of the mafic minerals separated from the same flows. The plagioclase crystals showed ideal Thellier behavior, with straight NRM/TRM trajectories.

6.3. Mid-Archean and Older Rocks

[82] Mid-Archean and older rocks ($\geq$3.0 Ga) have generally undergone at least low-grade metamorphism [e.g., *Xie et al.*, 1997; *Tice et al.*, 2004]; the time interval during which they experienced elevated temperatures could have spanned millions of years. In some cases this has led to the formation of new magnetic minerals [*Hale*, 1987]. These minerals can record stable magnetizations, and a nominal paleointensity can be derived, but the age of magnetization is uncertain. The absolute intensity is also uncertain because the efficiency of the TCRM process is unknown [*Dunlop and Özdemir*, 1997; *Yoshihara and Hamano*, 2004; *Smirnov and Tarduno*, 2005]. More generally, the low-grade metamorphism now places even tighter restrictions on the selection of samples for directional analysis or Thellier paleointensity study.

[83] Following *Néel*'s [1949, 1955] single-domain theory, the thermal relaxation time (τ) for single-domain grains can be expressed as [*Dunlop and Özdemir*, 1997]:

$$\frac{1}{\tau} = \frac{1}{\tau_0} \exp\left[-\frac{\mu_0 V M_s H_K}{2kT}\left(1 - \frac{|H_0|}{H_K}\right)^2\right], \qquad (6)$$

where τ_0 ($\sim 10^{-9}$ s) is the interval between thermal excitations, μ_0 is the permeability of free space, V is grain volume, M_s is spontaneous magnetization, H_K is the microscopic coercive force, k is Boltzmann's constant, T is temperature, and H_0 is the applied field. The microcoercive force provides a measure of the field needed for magnetization rotation in the absence of thermal excitation [*Dunlop and Özdemir*, 1997].

[84] This theory also can be used to derive time-temperature relationships that are useful for predicting the acquisition of thermoviscous magnetization. Given $H_K \gg H_0$ and relaxation times (τ_A and τ_B) representing two temperatures (T_A and T_B, respectively), we can write [*Pullaiah et al.*, 1975]:

$$\frac{T_A \ln(\tau_A/\tau_0)}{M_s(T_A)H_K(T_A)} = \frac{T_B \ln(\tau_B/\tau_0)}{M_s(T_B)H_K(T_B)}. \tag{7}$$

[85] This relation suggests that for the low-grade metamorphism discussed above (and a nominal reheating duration of 1 Myr), only single-domain grains with blocking temperatures greater than 400°C should be taken as firm evidence for the retention of a primary thermoremanent magnetization [*Dunlop and Buchan*, 1977].

[86] Whole rock samples are ensembles of multidomain, pseudosingle-domain, single-domain, and superparamagnetic grains. For multidomain grains we can expect the acquisition of partial thermoviscous magnetizations up to the Curie point of the magnetic mineral in question because of reequilibration of domain walls; this theory has been confirmed with detailed experiments [*Dunlop and Özdemir*, 1997]. With elevated temperatures, there is also the possibility of the acquisition of crystallization remanent magnetizations, related to the growth of new magnetic minerals or the alteration of primary magnetic minerals.

[87] Again, silicates that carry single-domain or pseudosingle grains provide an ideal alternative to whole rock measurement because the latter often contain multidomain magnetic grains which, as outlined above, are expected to carry magnetic overprints. However, what guides our choice of silicate when studying Phanerozoic or Proterozoic rocks might lead to a poor choice for investigations of the Archean. Given that thermally driven alteration is paramount, silicates with relatively high iron content might not be good choices because of the possibility of ionic diffusion and the formation of new magnetic minerals. Also, some silicates, particularly plagioclase feldspars, show physical alteration (i.e., saussuritization) under low-grade metamorphic conditions.

[88] Therefore the counterintuitive approach of choosing those silicates with the lowest intrinsic iron content may, in fact, be a way of learning about the Archean field. In this regard, measurements on quartz and microcline have already shown some promise. The relatively large crystal size of such silicates in plutonic rocks makes orientation feasible, opening the road to simultaneous determination of paleomagnetic directions and paleointensity [*Bauch et al.*, 2004]. Regardless of this choice, continued diligence (especially through optical scanning electron microscopy (SEM) and TEM analyses) is needed to insure that the silicate has not been compromised by later growth of magnetic minerals [e.g., *Otofuji et al.*, 2000].

[89] Other silicate grains hold promise for investigations of the early Earth's field. One of these is zircon; an appeal of this recorder is its tie to age, as zircons are typically analyzed for high-resolution Pb-Pb radiometric age data in Archean rocks. In fact, Franz magnetic separators are routinely used to identify (and exclude) zircons for such geochronology studies, providing some indication that zircons could contain inclusions of magnetic particles. *Lewis and Senftle* [1966] emphasized surface magnetic properties of zircons; it is not yet clear whether these properties, which can be removed by acid treatments, should be considered as a signal. Studies of zircons cleaned of surficial iron particles are at a nascent stage. However, the small size of simple zircons (often 0.1 mm) represents a challenge for measurement, even using the current generation of three-component DC SQUID magnetometers [*Libman-Roshal et al.*, 2000].

6.4. Exsolution in Plutonic Rocks

[90] As is the case for the Phanerozoic, silicates with magnetic inclusions incorporated during formation of the silicate crystal are ideal as this process may result in low anisotropy. However, for some Proterozoic and Archean intervals such rocks may not be available. Plutonic rocks with exsolved magnetic particles formed by slow cooling are relatively common. These can be used to study the field if factors related to anisotropy, slow cooling, and TCRM acquisition are systematically addressed.

[91] *Borradaile* [1994] argued that crystals with considerable magnetic fabrics could record paleofield directions if the fabric was oblate. *Selkin et al.* [2000] developed a paleointensity correction based on the measurement of TRM anisotropy. Although their method is directed toward whole rock samples, in principle, it can be applied to single silicate crystals. *Gee et al.* [2004] also describe methods of analyzing silicate fabrics and their uncertainties based on orthogonal sections.

[92] In slowly cooled rocks the differences between natural and laboratory cooling times are too large to ignore. Theoretical considerations to date suggest cooling rate corrections that are a function of domain state. With cooling on million-year timescales, raw data from Thellier analyses of single-domain grains will lead to strength overestimates approaching 50% [*Halgedahl et al.*, 1980]. Multidomain grains will underestimate the field intensity [*McClelland-Brown*, 1984]. TEM, SEM, and rock magnetic studies can be used to constrain grain sizes and shapes, and detailed radiometric dating studies provide a means to constrain cooling rates [*Briden et al.*, 1993]. However, further laboratory work to investigate the relationships between magnetic grain characteristics, cooling rate, and field recording are also needed.

[93] The uncertainties posed by TCRM are complex; they are relevant for both whole rock and single silicate samples. In whole rock diabase dikes, oxyexsolution of titanomagnetite can result in the formation of magnetite and ilmenite intergrowths [*Haggerty*, 1991]. If the process continues at temperatures below the Curie temperature of magnetite, as has been observed in some modern lava lakes [*Grommé et*

al., 1969], the resulting remanence is a TCRM. In single silicate grains, ionic diffusion is enhanced at high temperature. However, if the formation of magnetic minerals continues below the Curie point, as seems likely for some slowly cooled plutonic units, the remanence will also be a TCRM (or partial TCRM), and its efficiency relative to TRM must be investigated.

[94] A detailed evaluation of TCRM should start with constraints on the temperature at which the magnetic minerals are exsolved during cooling. For single silicate crystals, there has been some important progress. *Bogue et al.* [1995] adopted the method of *Fleet et al.* [1980] to constrain exsolution temperatures of opaques within clinopyroxene of the Duke Island ultramafic complex (Alaska). This method utilizes the measurable angles between the inclusion and host grain, angles which should reflect minimum strain. These angles can preserve information on the exsolution temperature because the two lattices expand differently when heated. *Bogue et al.* [1995] derived exsolution temperatures of 520°–550°C. *Feinberg et al.* [2004] used electron backscatter diffraction to constrain orientations and derived an exsolution temperature of approximately 840°C for magnetite exsolution in clinopyroxene from gabbros of the Messum Complex (Namibia); previous investigations had indicated that this exsolved magnetite was the source of stable remanence in the rocks [*Renne et al.*, 2002]. The estimate of exsolution temperature agreed well with that of approximately 865°C obtained from the amphibole-plagioclase geothermometer of *Holland and Blundy* [1994].

[95] From these studies it appeared that exsolution had occurred in some silicate crystals at temperatures sufficiently high such that the magnetization could be considered a TRM. However, further work by *Feinberg et al.* [2005] on the Messum Complex rocks indicates that the macroscopic pattern of exsolution does not provide sufficient information. Specifically, *Feinberg et al.* [2005] demonstrated that the needles described in prior studies were in some cases a "boxwork" of interacting single-domain magnetite particles separated by ulvospinel lamellae, thought to have formed from an unmixing of original titanomagnetite exsolved inclusions. The unmixing and resulting boxwork structure results in an increase in magnetic coercivity, but for our purposes this stability comes at a price: Because the boxwork likely formed at temperatures below the Curie point of magnetite, the remanence is a TCRM, and we lose the tie with the well-defined physics of TRM.

[96] Not all exsolved particles have the boxcar structure [*Frandsen et al.*, 2004; *Feinberg et al.*, 2005], and some may record a true TRM. However, in many other instances the temperatures of exsolution within a silicate host, or of unmixing of exsolved particles, are lower than the relevant Curie points, and the magnetization is a TCRM. TCRM is a special case of CRM. The CRM is blocked when a grain volume reaches a stable blocking volume V_B [e.g., *Kobayashi*, 1962] for a given temperature T:

$$V_B = \frac{2kT\ln(2t/\tau_o)}{\mu_o M_s H_K}, \tag{8}$$

where M_s is the spontaneous magnetization, μ_o is the permeability of free space ($4\pi \times 10^{-7}$ H/m), H_K is the microcoercivity of a grain, k is Boltzmann's constant, t is the characteristic time of the experiment, and τ_o is the atomic reorganization time.

[97] Assuming that the final grain size is identical and the characteristic relaxation times for TRM and CRM are comparable, *Stacey and Banerjee* [1974] obtained an expression for CRM efficiency:

$$\frac{\mathrm{CRM}}{\mathrm{TRM}} = \frac{H_K(T_B)}{H_K(T)} = \frac{K(T_B)}{K(T)}\left(\frac{M_s(T)}{M_s(T_B)}\right), \tag{9}$$

where T is the temperature at which CRM is acquired, T_B is the blocking temperature of the magnetic grains, and K is an anisotropy constant. Equation (9) predicts that CRM should always be less than TRM. The prediction has been supported by experimental evidence [*Stokking and Tauxe*, 1990] and further grain growth calculations [*McClelland*, 1996]. This implies that paleointensity data from some Archean-Proterozoic dikes in which high-temperature oxidation is common [*Hodych*, 1996], such as those from dikes of the Superior Province [*Macouin et al.*, 2003; *Halls et al.*, 2004], could underestimate the true field strength if the oxyexsolution continued below 580°C [*Smirnov and Tarduno*, 2005]. Equation (9) also indicates that paleointensity data reported on the oldest rocks to date, the ~3.5 Ga komatiites of South Africa, provide only a minimum field strength value because these serpentinized rocks clearly carry a CRM or TCRM [*Hale*, 1987; *Yoshihara and Hamano*, 2004]. Understanding TCRM to the point at which it can be used for paleointensity studies may ultimately require a return to study grain growth in the laboratory, possibly leading toward an empirically based calibration.

[98] Tools to identify TCRM are rapidly expanding. Some early aspirations for electron holography [e.g., *Tonomura*, 1987] are realized by applications which afford a view of nanoscale particles and their magnetic interactions [e.g., *Harrison et al.*, 2002]. Magnetic force microscopy [e.g., *Frandsen et al.*, 2004] can be used to define internal domain structures in exsolved particles. Developments in numerical simulations, or micromagnetic models, of fine-grained particles [e.g., *Williams and Wright*, 1998] can provide further insight into the nature of exsolved magnetic particles.

7. DISCUSSION AND SUMMARY

[99] Rock magnetic analyses, scanning electron microscopy, and transmission electron microscopy indicate that single silicate crystals can have fine-grained magnetic inclusions capable of recording the direction and intensity of the past magnetic field. Choices of rock and silicate-crystal type for study are important, because these choices are directly linked to magnetization process and magnetic inclusion domain state. To select a rock, one can rely on guidelines stemming from our understanding of how igne-

ous rocks and magnetic minerals form. For example, basaltic lavas are favored over felsic lavas because the latter cool from magmas that are close to the Curie temperatures of important magnetic mineral inclusions. The selection of silicate-crystal type, however, must reply on empirical data. Fortunately, it is now relatively straightforward to collect the critical discerning data prior to paleomagnetic analysis. For example, in many plutonic rocks magnetic hysteresis data indicate that feldspars should be favored over hornblende because the latter contains multidomain magnetic inclusions.

[100] Plagioclase feldspar separated from lavas contain minute magnetic inclusions which appear to have been protected from weathering. Thellier analyses using such crystals have been affirmed using an historic lava [*Cottrell and Tarduno*, 1999]. Subsequent tests show that plagioclase crystals are less susceptible to experimental alterations than whole rocks [*Cottrell and Tarduno*, 2000]. Because the plagioclase feldspars can be collected from long basalt sequences, they afford a means to obtain joint paleointensity and secular variation data.

7.1. Phanerozoic Geomagnetic Field

[101] Thellier results from whole rock lava samples have led some authors to conclude that geomagnetic reversal rate and paleointensity are decoupled [e.g., *Prévot et al.*, 1985]. There are several features of the whole rock data that give reason for pause. Only a few paleointensity results from rocks older than 10 million years are based on multiple, independent cooling units that span significant secular variation. Moreover, there is a large difference between the mean field value from lavas older and younger than 10 Ma; the older rock suggests low field values [*Tarduno and Smirnov*, 2004]. These low values could reflect the presence of nonideal magnetic carriers, laboratory alteration, and maghemitization. Weathering results in nonreversible changes to a whole rock matrix and magnetic mineral phenocrysts, and these effects are ubiquitous in older lavas. The formation of clays in a whole rock matrix essentially primes rocks for experimental alteration. Magnetic mineral phenocrysts are transformed by low-temperature oxidation and maghemitization, replacing the certainty of TRM with the ambiguity of CRM. Some of these effects lead to obvious failures of experimental reliability tests. Others are more subtle and can contribute to a bias toward underestimates of field strength.

[102] Because of these limitations, it is inappropriate to apply statistical treatments to the entire set of virtual dipole moments derived from whole rock lava samples [e.g., *Heller et al.*, 2002; *Biggin and Thomas*, 2003] to draw conclusions on the geodynamo. Although there are some rigorous Thellier data sets from older lavas, natural and attendant laboratory alteration will fundamentally limit progress in understanding long-term paleointensity history using bulk samples. This is a prime motivation for a continued pursuit of single-silicate-based Thellier analyses.

[103] In particular, this approach provides the opportunity to investigate relationships between the frequency of geomagnetic reversals and the morphology, secular variation, and intensity of Earth's magnetic field. These relationships should be best expressed during superchrons. Thellier analyses of plagioclase crystals from lavas of the 113–116 Ma Rajmahal Traps of eastern India and ~95 million year old lavas of the Arctic indicate that the time-averaged field was remarkably strong ($>12 \times 10^{22}$ Am^2) during these portions of the Cretaceous Normal Polarity Superchron [*Tarduno et al.*, 2001, 2002a]. The field was also overwhelmingly dipolar, lacking significant octupole components [*Tarduno et al.*, 2002a]. These data suggest that the basic features of the geomagnetic field are intrinsically related. Superchrons may reflect times when the nature of core-mantle boundary heat flux allows the geodynamo to operate at peak efficiency, as suggested by some numerical dynamo simulations [e.g., *Glatzmaier et al.*, 1999].

[104] Thellier paleointensity analyses of plagioclase crystals from an interval of moderate reversal frequency during the Paleogene, and a period of high reversal occurrence during the Late Jurassic, further support an inverse relationship between reversal rate and intensity [*Tarduno and Cottrell*, 2005]. The data available to date also suggest that the strength of the time-averaged reversing field is more variable than that of the nonreversing field. These observations are consistent with an active lower mantle, shaping core-mantle boundary heat flux and efficiency of the geodynamo on timescales of tens of millions of years.

[105] Thellier analyses of plagioclase separated from lavas can be used to further test the relationships described above and the overall history of the geodynamo during the Phanerozoic. An obvious target is the Kiaman Reversed Polarity Superchron and intervals bounding it. Even older Paleozoic intervals are intriguing, but it should be stressed that the basic outline of field behavior provided by magnetostratigraphy and secular variation studies is still a work in progress.

7.2. Magnetic Field of the Young Earth, the Moon, and Other Planets

[106] A frontier area is the Precambrian and the long-term history of the geomagnetic field bracketing inner core growth. For Proterozoic-Archean times, dikes and some lavas are available for study. However, nearly all bulk samples of these rocks contain nonideal carriers in the form of multidomain magnetic grains. Multidomain inclusions in the mafic minerals can contribute to underestimates of the field if high-temperature segments of NRM/TRM data are used to calculate paleointensity values [*Dunlop et al.*, 2005]. An alternative approach, utilizing rigorous Thellier analyses applied to single silicate grains, holds significant promise for revealing this history, but rocks must be carefully chosen to minimize uncertainties. As is the case for the Phanerozoic, a focus on silicates with inclusions which together have minimal anisotropy are desirable. Significant promise has already been demonstrated through the study of single plagioclase crystals separated from Proterozoic-Archean dikes of Karelia [*Smirnov et al.*, 2003]. While the results are too few to average secular

variations, the field strengths are comparable to those of the modern field. Paleomagnetic directional data from Karelia and other sites [*Smirnov and Tarduno*, 2004] suggest a dipole-dominated field that is also similar to the recent field, and possibly more dipolar, consistent with some numerical models of the geodynamo with a smaller solid inner core [*Roberts and Glatzmaier*, 2001].

[107] A prominent concern for the Archean is the potential for later magnetic grain growth associated with low-grade metamorphic events. Silicates with low interstitial Fe, and hence a lower potential for the growth of magnetic minerals under low-grade metamorphic conditions, may be the best paleomagnetic targets. For some time slices of the Precambrian, however, ideal rocks will not be available. Rocks with stable magnetic signals carried by exsolved magnetic mineral in silicates are more common. The uncertainties posed by anisotropy, cooling rate, and thermochemical remanent magnetization (TCRM) are large but perhaps not insurmountable. Because of the relationship between crystallographic planes and Fe-oxide growth, anisotropy is an expected characteristic of magnetic mineral exsolution in silicates. Corrections can be sought using thermoremanent magnetization anisotropy [*Selkin et al.*, 2000]. Cooling rate corrections can be developed using nature cooling curves constrained by radiometric data and detailed magnetic grain size information constrained by rock magnetism, SEM, and TEM studies [e.g., *Briden et al.*, 1993]. The possibility of TCRM can be investigated by the relationship between magnetic inclusions and host grains in some silicates [e.g., *Bogue et al.*, 1995] and by advanced microscopy (TEM and magnetic force [e.g., *Feinberg et al.*, 2005]). If the magnetization is a TCRM, its efficiency relative to a TRM must addressed before absolute paleointensity values can be claimed [*Smirnov and Tarduno*, 2005].

[108] Many of the challenges of TCRM may apply to efforts to understand the magnetic signature of rocks from the Moon, Mars, and other bodies in the solar system. As in the case of terrestrial rocks the study of single silicate crystals may be a better approach to learning about the past magnetic field environment. It is also important to consider the scope of the problem of interest. For some of the oldest rocks on Earth the definition of directions using oriented single plagioclase crystals provides a means to see through the veil of low-grade metamorphism, revealing geomagnetic field behavior. Documentation of the absence/presence of a dipolar field may in itself be a significant contribution on Earth and on other planets. These are demanding endeavors, but our rapidly expanding ability to read magnetic history using single silicate crystals puts many of the answers within reach.

[109] **ACKNOWLEDGMENTS.** Our fieldwork in the high Arctic involved paleomagnetic sampling in sometimes inhospitable conditions. We thank O. Libman-Rosahal, H. Scher, J. Totten, and K. Weaver, who assisted in these collections while undergraduates at the University of Rochester. Logistic support for studies in the Arctic came from the Canadian Polar Shelf Project. We are grateful to P. K. Bandyopadhyay, S. S. Sarkar, and M. Mukul for assistance in field studies in northeastern India. We thank P. Lippert for his help in field studies of the Matachewan dikes and R. Coe for providing samples from Hawaii. We thank B. McIntyre for help with SEM and TEM analyses at the University of Rochester and the University of Minnesota's Institute for Rock Magnetism (funded by NSF and the Keck Foundation) for assistance in obtaining low-temperature data. We also are grateful to A. Kosterov for providing low-temperature data on single silicate crystals and Michael Watkeys for his assistance in field studies in southern Africa. This research used samples provided by the Ocean Drilling Program, which was sponsored by the U.S. National Science Foundation (NSF) and participating countries under management of the Joint Oceanographic Institutions. We thank the ODP Leg 197 Shipboard Scientific Party for helpful discussions and R. Coe, M. Fuller, and an anonymous reviewer for their helpful comments. Overall support for the authors' work came from NSF.

[110] The Editor responsible for this paper was Michael Manga. He thanks technical reviewers Robert Coe and Michael Fuller and one anonymous cross-disciplinary reviewer.

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