SPECIAL TOPICS IN ROCK MAGNETISM

In Chapter 3, you discovered the basic mechanisms by which NRM is formed. A variety of special topics in rock magnetism are investigated in this chapter. These topics include (1) special attributes of some types of NRM, such as the ability to retrieve paleointensity of the geomagnetic field from TRM; (2) concerns about effects of chemical changes on primary NRM; (3) questions about the accuracy of NRM records, in particular the questions of inclination error in DRM and possible directional effects of magnetic anisotropy; and (4) the timing of remanence acquisition by red sediments. There are no definitive answers to some of these questions and concerns. But consideration of these topics is important to interpretation of paleomagnetic data in the coming chapters.

PALEOINTENSITY FROM THERMOREMANENT MAGNETIZATION

The development of thermoremanent magnetism in Chapter 3 focused on directional properties of TRM. But TRM is unique among forms of natural remanent magnetism in providing information about past intensities of the geomagnetic field via a technique that is straightforward in principle. Consider Equation (3.28), which describes dependence of TRM on various parameters including strength of magnetizing field, *H*:

$$\mathrm{TRM}(20°\mathrm{C}) = N(T_B)\, v\, j_s(20°\mathrm{C})\, \tanh(b) \tag{8.1}$$

where

$$b = \frac{v\, j_s(T_B) H}{k\, T_B}$$

For typical values of parameters, b is << 1.0. This provides a useful simplification because tanh (b) ≈ b for b << 1.0. Thus from Equation (8.1),

$$\mathrm{TRM}(20°\mathrm{C}) \approx N(T_B)\; v\, j_s(20°\mathrm{C}) \left(\frac{v j_s[T_B] H}{k T_B} \right) \tag{8.2}$$

TRM thus depends <u>linearly</u> on the strength of the magnetic field present during cooling through the blocking temperature. The magnetic field dependence can be made more explicit by combining terms that depend on grain-size and shape distribution, blocking temperature, and ferromagnetic properties (e.g., $N(T_B)$, $j_s(T_B)$, etc.) into a proportionality constant, A. Equation (8.2) becomes

$$\mathrm{TRM}(20°\mathrm{C}) = \mathrm{A}H \tag{8.3}$$

If the TRM under consideration formed by cooling in the geomagnetic field, this natural TRM (TRM_{paleo}) is linearly dependent on the intensity of the paleomagnetic field, usually referred to as *paleointensity*. The paleointensity experiment is designed to determine the proportionality constant, A.

Suppose you are attempting to determine the paleointensity of the geomagnetic field from a particular rock sample that contains a primary TRM that we call TRM_{paleo}. It is an easy matter to measure TRM_{paleo}, but neither term on the right side of Equation (8.3) is known. In principal, the proportionality constant A can be determined by giving the same sample a new TRM (TRM_{lab}) in a known field, H_{lab}, so that

$$TRM_{lab} = A\, H_{lab} \tag{8.4}$$

If the natural TRM, TRM_{paleo}, is an uncomplicated, single-component TRM, the paleointensity can be obtained by combining Equations (8.3) and (8.4) to eliminate the proportionality constant, A. Solving for H_{paleo} yields

$$H_{paleo} = \left(\frac{\mathrm{TRM_{paleo}}}{\mathrm{TRM_{lab}}}\right) H_{lab} \tag{8.5}$$

In principle, all quantities in Equation (8.5) are easily measurable, and paleointensity can be determined. However, the entire experiment depends on the assumption that no changes occur in the proportionality constant, A. This means that no changes in any properties that determine A (grain size or composition of ferromagnetic grains, etc.) can take place in nature since the original TRM formed <u>or</u> during laboratory heating. In practice, paleointensities of the geomagnetic field are very difficult to determine because the ferromagnetic grains carrying the natural TRM have often undergone alteration and/or the required laboratory heating induces physical or chemical changes. To extract useful paleointensity information at low temperatures before higher temperature alterations occur, the paleointensity experiment is usually done in a series of heating steps to progressively higher temperatures.

The procedure involves a double-heating process:

1. The sample is first heated to a temperature T_i above room temperature but below the Curie temperature. The sample is then cooled to room temperature in zero magnetic field, and the TRM_{paleo} remaining in the sample is measured. The difference between TRM_{paleo} prior to heating and TRM_{paleo} after heating to T_i is the amount of natural TRM with blocking temperatures $\leq T_i$; this difference is the natural partial thermoremanent magnetism (PTRM) carried by grains with blocking temperature $\leq T_i$.

2. Again the sample is heated to T_i but is cooled in a known magnetic field H_{lab}. The amount of PTRM acquired during this cooling is then measured. The TRM_{paleo} remaining after the first heating to T_i is plotted against the PTRM acquired by cooling in H_{lab} following the second heating. This double-heating process is repeated at incrementally higher temperatures; a data point plotting remaining TRM_{paleo} against acquired PTRM is obtained at each temperature. An example plot is shown in Figure 8.1.

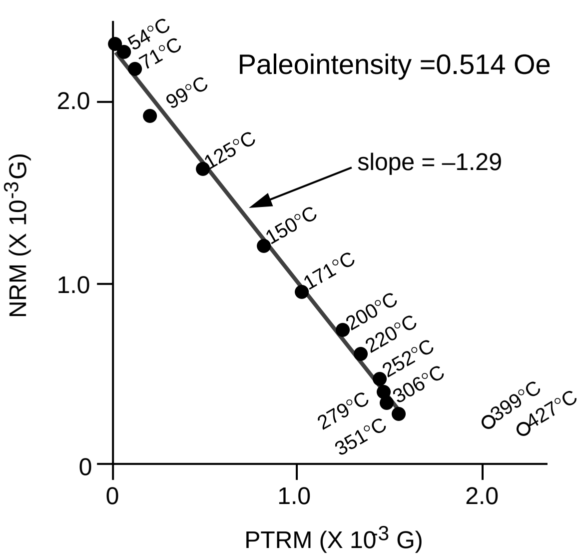

Figure 8.1 NRM remaining versus PTRM acquired. Data points plot NRM remaining after heating to a particular temperature against PTRM acquired by heating to the same temperature followed by cooling in a 0.4-Oe magnetic field; temperatures of heating are shown adjacent to data points; the slope of the line fit to the data points is –1.29; the sample is 3790-year-old basalt from Hawaii, and NRM is a primary TRM. Redrawn from Coe et al. (1978), with permission of the American Geophysical Union.

The law of additivity of PTRM (Equation (3.30)) says the PTRM acquired in one interval of blocking temperature is independent of PTRM acquired in other intervals. So Equation (8.5) can be applied at each temperature T_i and each data point on a paleointensity plot (Figure 8.1) provides an estimate of the paleointensity H_{paleo}. If no changes occur in the ferromagnetic grains, the data points obtained at progressively higher temperatures fall on a straight line. The slope of this line is

$$slope = -\frac{H_{paleo}}{H_{lab}} \tag{8.6}$$

In the example of Figure 8.1, the slope is –1.29, the laboratory field used was 0.4 Oe, and the resulting paleointensity is 0.514 Oe.

If heating above a certain temperature causes changes in the ferromagnetic grains, data points from higher temperatures will not fall on the line described by the data obtained at lower temperatures. Note that data points obtained at 399°C and 427°C in the example of Figure 8.1 fall off the line described by the lower temperature data. This indicates that changes affecting the ferromagnetic minerals have taken place when the sample was heated to >350°C. Only the lower temperature data should be used to determine the paleointensity.

By employing this double-heating procedure, useful paleointensity information can often be obtained at low temperatures before higher-temperature alteration occurs. But the procedure is time consuming, and the success rate is sometimes low. As a consequence, much more is known about past directions of the geomagnetic field than about past intensities. However, knowledge of geomagnetic paleointensities is crucial to evaluation of geomagnetic field models. Accordingly, much effort has been put into development and use of paleointensity techniques.

Merrill and McElhinny (1983) provide a more thorough discussion of paleointensity experiments and results. The book edited by Creer et al. (1983) contains a number of articles on paleointensities. Figure 1.10 showing variations in the geomagnetic dipole moment over the past 10^4 yr was determined from paleointensity experiments compiled by McElhinny and Senanayake (1982). For a discussion of field intensity during a reversal of the geomagnetic dipole, see Prévot et al. (1985).

INCLINATION ERROR OF DRM

In Chapter 3, inclination error (inclination shallowing) of detrital remanent magnetism was discussed in the context of DRM acquisition. Here we investigate how and when inclination error may develop and its likely magnitude in various sedimentary environments. An obvious question is: *Does this inclination error of DRM happen in nature?* The definitive answer is: *Probably, sometimes*.

We have already observed (Figure 3.15) that inclination of depositional DRM, I_o, is systematically shallower than inclination of the magnetic field at the time of deposition, I_H. In a number of redeposition experiments (King, 1955; Griffiths et al. 1960; King and Rees, 1966), the inclinations were found to be related by

$$\tan I_o = f \tan I_H \tag{8.7}$$

and the value of f was found to be ~0.4 for redeposited glacial sediments. The corresponding inclination error, ΔI, is

$$\Delta I = I_H - I_o = I_H - \tan^{-1}(f \tan I_H) \tag{8.8}$$

But postdepositional DRM (pDRM) processes dominate magnetization of many sediments, especially fine-grained sediments. And pDRM produces accurate recordings of the direction of the magnetic field (Irving and Major, 1964; Opdyke and Henry, 1969; Kent, 1973; Barton and McElhinny, 1979).

Two natural examples often cited as evidence for absence of inclination error in pDRM are now discussed:

1. *Paleomagnetic records from Holocene lake sediments.* Although exceptions exist, high-quality paleomagnetic records from Holocene lake sediments usually record the inclination of the geomagnetic field at or soon after deposition. The evidence is convincing: (a) many lake sediment paleomagnetic records agree with historic geomagnetic field records; (b) other lake sediment paleomagnetic records agree with directions of thermoremanent magnetism recorded by archeological features or Holocene lava flows; (c) mean inclination observed in sequences of lake sediments spanning >10^3 yr usually agree with expected inclination of a geocentric axial dipole (Lund, 1985).
2. *Paleomagnetic records from Plio-Pleistocene deep-sea cores.* Opdyke and Henry (1969) examined paleomagnetism of piston cores of deep-sea sediments collected from a wide variety of locations. These cores allowed collection of only the upper few meters of sediment, which is usually no older than Early Pliocene (ca. 5 Ma). Mean paleomagnetic inclinations are plotted against latitude of collecting site in Figure 8.2; the curve in the diagram is the expected inclination for a geocentric axial dipole field. Fundamental agreement of the observed mean inclination with the predicted inclination argues that no inclination error exceeding about 5° is present.

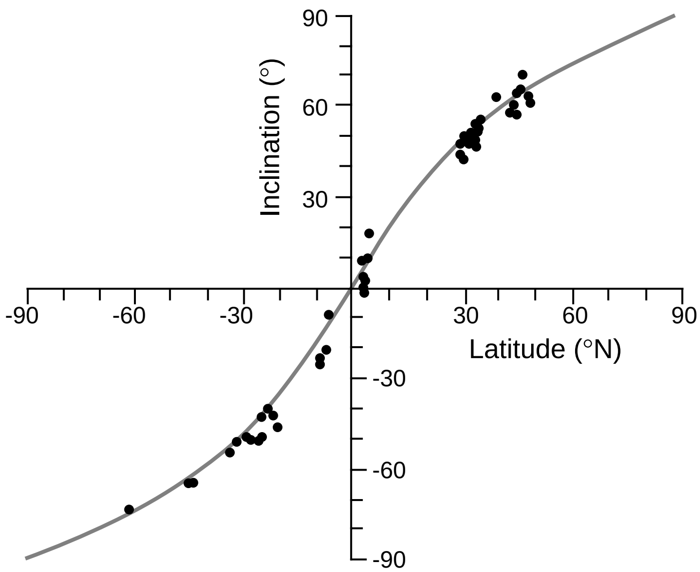

Figure 8.2 Mean inclination of DRM in 52 Plio-Pleistocene deep-sea sediment cores versus latitude of core collection. The solid curve shows the expected inclination for the geocentric axial dipole field; to calculate the mean inclination for each core, the sign of the inclination of DRM in reversed-polarity intervals was changed. Redrawn from Opdyke and Henry (1969).

These examples demonstrate that fine-grained sediments with magnetization dominated by pDRM processes and buried by a few meters of overlying sediments do not possess inclination error. But these results do not demonstrate lack of inclination error in older sediments. Recent examinations indicate that compaction (and possibly deformation) can often shallow the inclination of magnetization.

The potential importance of compaction can be understood by considering changes in porosity resulting from compaction. Clays have typical initial porosity of 50% to 80%. Porosity decreases by about 50% on burial to 1 km; almost complete closure of pores occurs on burial to 2 km. Sands have initial porosity of 20% to 65%, and burial to 4 km decreases porosity to about one half of the initial value. These porosity changes demonstrate the potential for compaction-induced rotation of platy and elongate grains toward the bedding plane. Shallowing of inclination could result in much the same manner shown in Figure 3.15b.

Deep-sea sediments that are older (and more deeply buried) than those of Figure 8.2 sometimes have inclinations of magnetization shallowed by compaction (Blow and Hamilton, 1978; Celaya and Clement, 1988; Arason and Levi, 1990a). Recent advances in ocean floor drilling have allowed the retrieval of cores several hundred meters in length. In some cores up to 500 m in length with oceanic sediments no older than Miocene, inclinations are seen to systematically shallow down core by some 10° to15° due to compaction. A corresponding gradual decrease in water content is observed, and clay particles are rotated toward the bedding plane. An example is given in Figure 8.3.

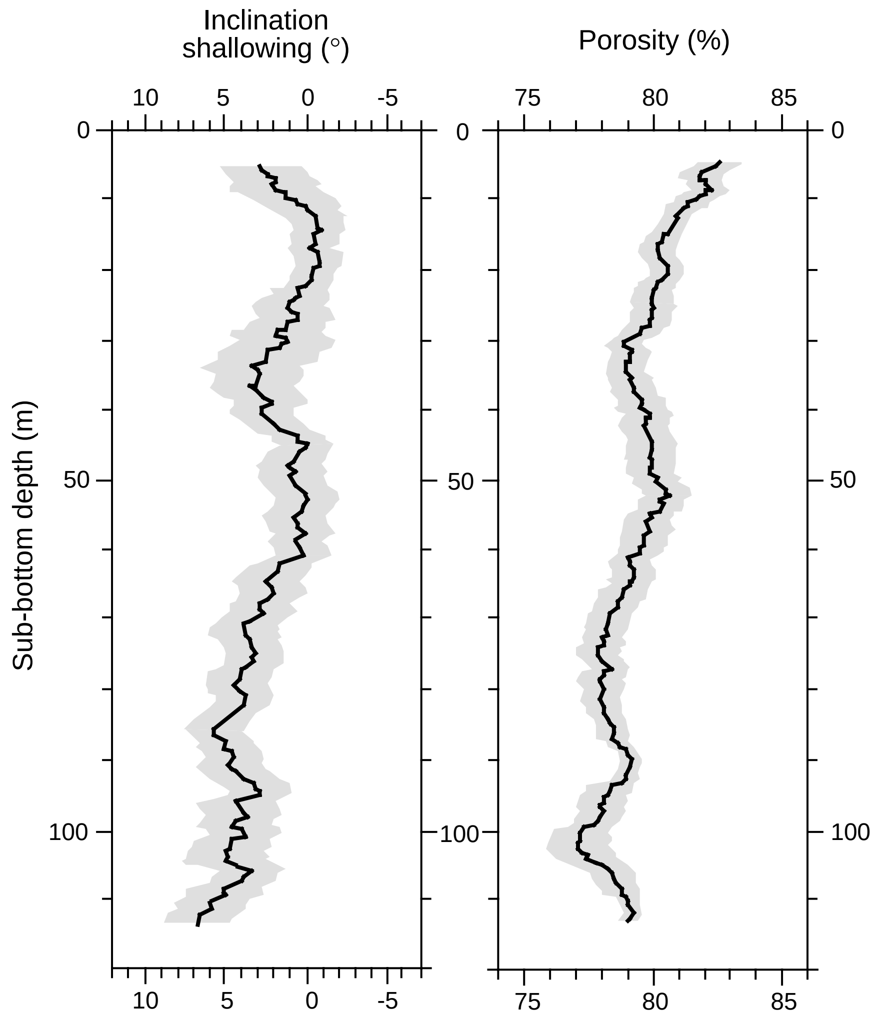

Figure 8.3 Shallowing of DRM inclination and porosity versus depth in a core of deep-sea sediment. The core is from DSDP Site 578 in the northwestern Pacific Ocean; the oldest sediment has an age of 5.6 Ma; the bold line is a 1-m.y. sliding time-window average of inclination shallowing compared with geocentric axial dipole field inclination (corrected for Pacific plate movement); porosities are means calculated every 0.2 m using a 10-m sliding-depth window; stippled envelopes show 95% confidence limits. Redrawn after Arason and Levi (1990), with permission of the American Geophysical Union.

Laboratory experiments suggest that interactions between fine-grained magnetite and clay particles may be important in compaction shallowing of inclination (Anson and Kodama, 1987; Deamer and Kodama, 1990). Small elongate magnetite particles are thought to adhere to clay particles or be trapped inside clusters of clay particles. During compaction, the long axes of magnetite grains are passively rotated toward the bedding plane along with the clay particles. Arason and Levi (1990b) have investigated a variety of models for compaction shallowing of inclination.

For older sedimentary rocks, evidence for or against inclination error becomes less clear. This evidence must come from comparison of paleomagnetic records from sedimentary and igneous rocks of identical age. Such comparisons are not simple because tectonic histories, adequate sampling of geomagnetic secular variation by the igneous rocks, and other complicating factors must be taken into account. Nevertheless, there are a few well-documented examples.

1. *Eocene turbidites of the Oregon Coast Range.* In the Oregon Coast Range, Eocene turbidites of the Tyee and Flournoy formations are overlain by the Tillamook Volcanic Series and underlain by

the Siletz River Volcanics. Both the volcanic rocks and the turbidites have been the subject of extensive paleomagnetic study (Simpson and Cox, 1977; Magill et al., 1981). The inclination of DRM in the Tyee and Flournoy formations closely matches inclinations in the bracketing volcanic sequences. This is clear evidence against significant inclination error in these turbidites.

2. *Alaskan terranes*. Paleomagnetic data from Late Cretaceous and Early Tertiary lavas and oceanic sedimentary rocks are available from several tectonostratigraphic terranes in Alaska. Comparing paleomagnetic data from sediments and lavas of the same terranes, Coe et al. (1985) found that sedimentary rocks yield systematically shallow inclinations. For the Prince William and Chugach terranes, paleomagnetic inclinations from sediments are about 20° shallower than inclinations from lavas. The value of f in Equations (8.7) and (8.8) that best describes the shallowed inclinations in the Alaskan turbidites is $f \approx 0.4$. Many of these sedimentary rocks are deformed, so that shallowed inclinations might have been produced by deformational effects as well as by compaction.

3. *Paleocene continental sediments of San Juan Basin, New Mexico*. Continental claystones and fine siltstones of this Laramide basin were the subject of extensive paleomagnetic study (Butler and Taylor, 1978). The Nacimiento Formation of Paleocene age yielded high-quality paleomagnetic data with many stratigraphic levels investigated. These data were used in Figure 5.16 as an example of the reversals test; means of the normal- and reversed-polarity sites are antipodal to within 1.6°. Yet the mean inclination is 8° ± 3° shallower than predicted by paleomagnetic poles determined from Paleocene igneous rocks in Montana. This shallowing of inclination is almost certainly the effect of compaction.

4. *Late Cretaceous and Early Tertiary pelagic limestones, Umbrian Apennines, Italy*. An extraordinary amount of paleomagnetic data are available from pelagic limestones of northern Italy (see Chapter 9). Inclination of magnetization in these limestones is indistinguishable from the expected inclination predicted for the African plate to which these limestones were formerly attached (Lowrie and Heller, 1982). No inclination error exists in the paleomagnetism of these pelagic limestones.

5. *Pacific Plate Deep Sea Drilling Project Sediments*. Many sediment cores have been collected from the Pacific Ocean Basin by the Deep Sea Drilling Project (DSDP). Paleomagnetic inclinations from these cores have been used to determine the paleolatitude at which these sediments were deposited (using Equation (1.15)). Paleolatitudes can be determined from a variety of other observations, including (1) paleomagnetic data from sequences of lava flows collected at some DSDP sites, (2) analysis of magnetic anomalies produced by seamounts, (3) analysis of the shape of lineated marine magnetic anomalies, and (4) sedimentologic determination of facies deposited near the equator. From all methods of analysis, it is clear that portions of the Pacific Plate moved into the northern hemisphere from Cretaceous paleolatitudes in the southern hemisphere. Tarduno (1990) and Gordon (1990) have shown that the southerly paleolatitudes determined from paleomagnetism of Pacific DSDP sediments are systematically lower (closer to the equator) than paleolatitudes determined from the other techniques.

A shallowing of the paleomagnetic inclination (Equation (8.7)), leads to an error in the paleolatitude (λ) determined from the mean inclination. This paleolatitude error, $\Delta\lambda$, is given by

$$\Delta\lambda = \lambda - \tan^{-1}(f \tan \lambda) \qquad (8.9)$$

where λ is the paleolatitude at which the sediments were deposited. For Pacific DSDP sediments, Tarduno (1990) found a best-fit value of $f = 0.52$ with lower and upper confidence limits of $f = 0.23$ and $f = 0.80$.

Thus, it appears that inclination error of about 10° can be documented for some sediments, whereas absence of inclination error can be demonstrated for other sedimentary rocks. We cannot yet predict which rock types contain inclination error. Nevertheless, we can make some generalizations about sources of inclination error and sedimentary rocks that are most likely to contain inclination error.

1. *Depositional inclination error*. Shallowed inclinations during acquisition of depositional DRM (Figure 3.15b) are most likely to occur in larger grain-size sediments. High deposition rate may enhance this effect. For most fine sands and smaller grain-size sediments and any bioturbated sediment, postdepositional alignment dominates and has the effect of erasing depositional inclination error.
2. *Compaction*. Shallowing of inclination can be induced by compaction and is probably a larger effect for fine-grained sediments. Lithologies that undergo substantial compaction (e.g., claystone, mudstone, or sediments with muddy matrix) are probably most susceptible to inclination shallowing through compaction. Lithologies showing minimal compaction such as grain-supported sandstones might not experience compaction shallowing of inclination.
3. *Deformation*. It is likely that deformation can affect inclination. Folding of sedimentary strata involves strain, and high degrees of strain might realign magnetic grains producing magnetic anisotropy. Inclination error might be a result.
4. *Cementation*. While there are many unknowns regarding inclination error, it is clear that early cementation prevents compaction-induced inclination error because cementation essentially halts compaction. Sedimentary rocks that have been cemented soon after deposition are probably immune to shallowing of DRM by compaction.

BIOMAGNETISM: BIRDS DO IT, BEES DO IT

Recent research indicates that magnetite is a biochemical precipitate of major significance. *Biogenic magnetite* has been found in three of the five kingdoms of living organisms, including pigeons and honeybees. Although originally thought to be unrelated to paleomagnetism, biogenic magnetite has been found in a wide variety of sedimentary rocks and might be a major contributor to DRM in marine sediments (Chang and Kirschvink, 1989).

The most celebrated examples of organisms containing biogenic magnetite are *magnetotactic bacteria*. These bacteria contain magnetite crystals arranged in chains and held within a *magnetosome*. Transmission electron microscopy (TEM) has revealed that magnetite grains in magnetotactic bacteria (and in a wide variety of other organisms) are within the stable single-domain (SD) size and shape range (Figure 8.4). Accordingly, individual SD magnetite crystals in bacterial magnetosomes have maximum intensity and stability of magnetization. Furthermore, magnetite crystals are aligned within the magnetosome so that magnetocrystalline easy directions are parallel to the chain with the result that magnetic moments of individual crystals add up to produce a very effective and stable magnet. This magnet serves as a geomagnetic sensor that guides magnetotactic bacteria down magnetic flux lines, helping them to remain within the preferred habitat of oxygen-poor zones within muddy layers of accumulating sediment.

TEM examinations have shown that biogenic magnetite crystals have morphologies that are distinct from magnetite of igneous or authigenic origin. An example is presented in Figure 8.5a. Examination of morphology of magnetite crystals therefore allows identification of biogenic magnetite in sedimentary rocks, and these magnetites are referred to as *magnetofossils*. Biogenic magnetite has been found in marine sedimentary rocks as old as 700 Ma from a wide variety of depositional environments and are especially prevalent in calcareous oozes. Estimates of bacterial abundances and sediment accumulation rates indicate that biogenic magnetite could account for a major percentage of stable DRM in marine sedimentary rocks. The paleomagnetic significance of biogenic magnetite is emphasized by the observation that all sedimentary rocks that are shown to contain biogenic magnetite also contain a stable paleomagnetism formed as a primary DRM.

MARINE SEDIMENTS

Marine sediments are a rich potential source of paleomagnetic data because biostratigraphic data can provide accurate age information and thick sections can encompass large time intervals. In addition, numerous

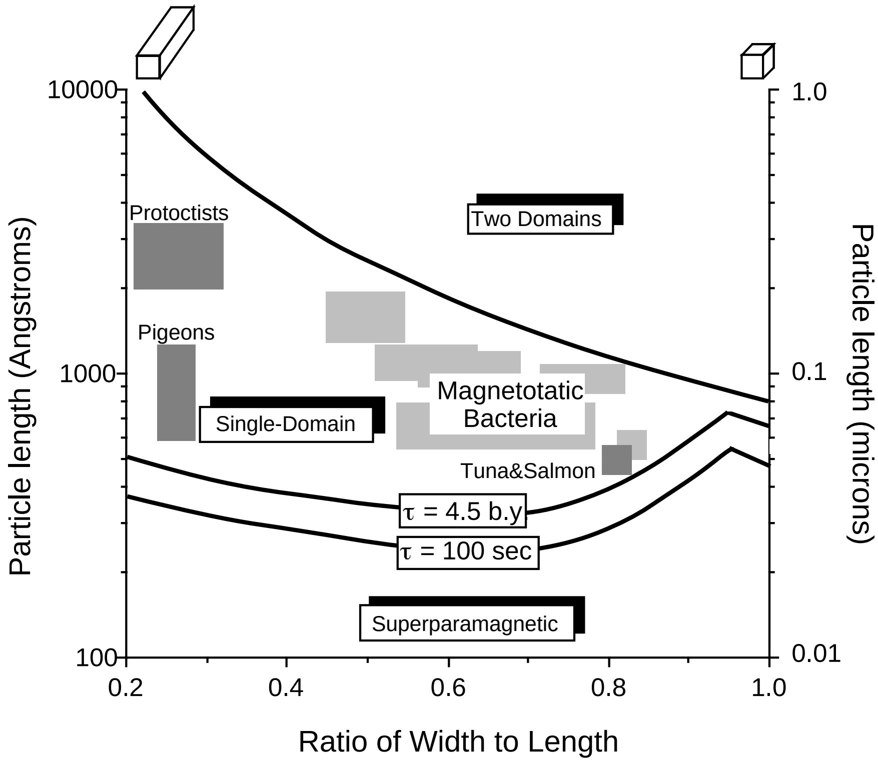

Figure 8.4 Size and shape distribution of biogenic magnetite grains. Distribution of grains in magnetotactic bacteria is shown by lightly stippled fields; distribution of grains in other organisms is shown by darker stippled fields; distribution of two-domain, single-domain, and superparamagnetic fields is from Figure 3.2. Redrawn from Chang and Kirschvink (1989). Reproduced, with permission, from the Annual Reviews of Earth and Planetary Sciences, Vol. 17, copyright 1989 by Annual Reviews Inc.

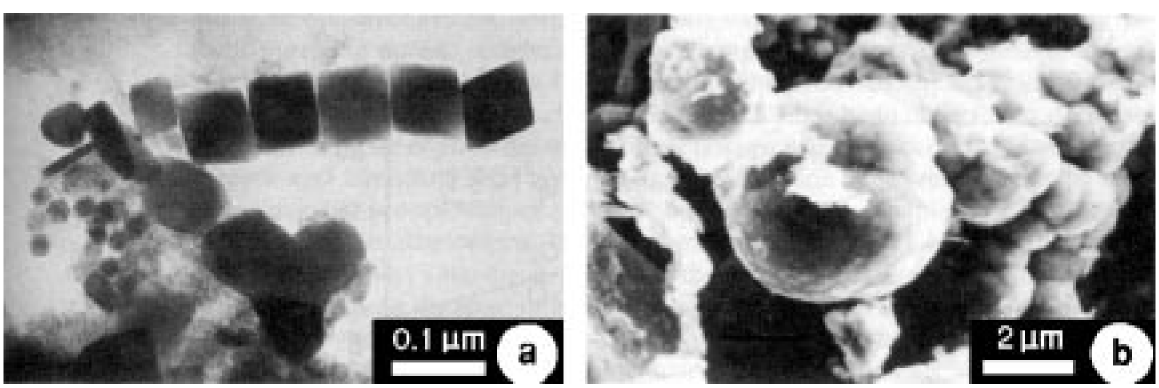

Figure 8.5 (**a**) Transmission electron micrograph of biogenic magnetite crystals from a deep-sea sediment. Kindly provided by H. Vali. (**b**) Scanning electron micrograph of botryoidal authigenic magnetite in the Helderberg Group (Devonian) of New York state. Kindly provided by C. McCabe.

subaerially exposed sections of marine sediments (especially shallow-water carbonates) are available. Although intensities of remanent magnetization are low (typically 10^{-6} to 10^{-8} G, 10^{-3} to 10^{-5} A/m), modern magnetometers can measure these weak magnetizations quite accurately.

Some deep-sea cores and subaerial sections of marine sediments yield high-quality paleomagnetic data, while others do not. Destruction of original detrital ferromagnetic minerals and late diagenetic production of ferromagnetic minerals are basic reasons for failure to obtain useful paleomagnetic data. In this section, we consider some fundamental geochemistry of marine sediments. For a more complete discussion, see the excellent review by Henshaw and Merrill (1980).

The first consideration is stability of iron oxides and sulfides in marine sedimentary environments. An equilibrium diagram for the Fe–S–H_2O system is shown in Figure 8.6. The small stippled box in the figure indicates the range of normal seawater conditions. The pH of seawater and marine sediments is controlled within a narrow range ($8.1 < pH < 8.2$). But oxidizing or reducing conditions vary widely from the nominally oxidizing conditions of seawater to highly reducing conditions within sediments containing abundant organic matter. Figure 8.6 shows that goethite is the Fe-oxide expected to precipitate from solution under normal conditions (if Fe exceeds solubility limits). However, authigenic magnetite and/or pyrite may precipitate if neutral or reducing conditions occur during diagenesis.

Hemipelagic sediments

Hemipelagic sediments have at least 25% of coarse fraction composed of terrigenous, volcanogenic, and/or neritic detritus. These sediments are usually deposited on the continental margin and adjacent abyssal plain. Rates of sediment accumulation are typically 1 m/1000 yr. The dominant detrital ferromagnetic mineral is magnetite with typical concentration 0.05% by volume. Grain size of magnetite is dominantly ≤1 μm. This magnetite is an efficient recorder of primary DRM.

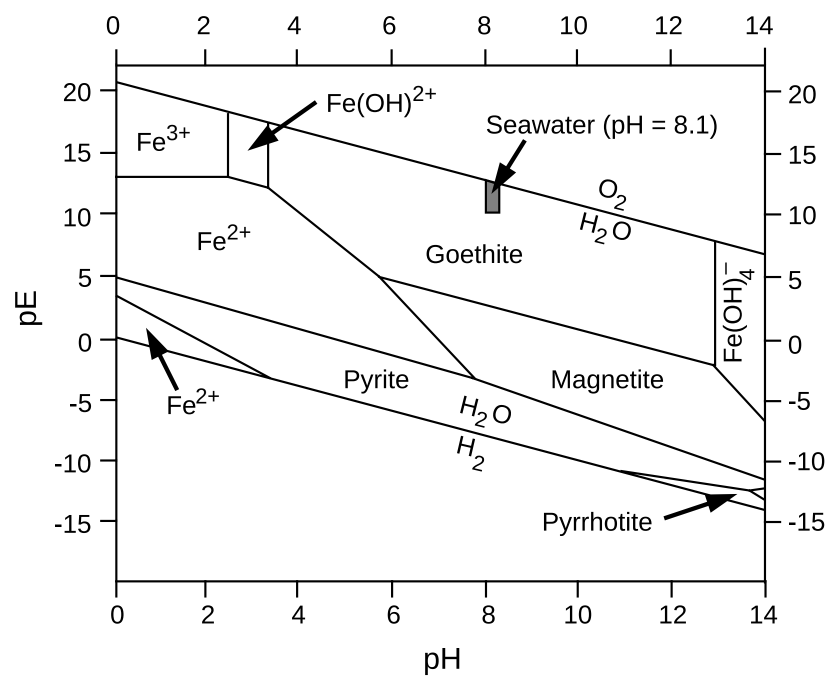

Figure 8.6 Equilibrium diagram of the Fe–S–H_2O system. $pH < 7$ indicates acidic conditions; $pH > 7$ indicates basic conditions; $pE > 0$ indicates oxidizing conditions; $pE < 0$ indicates reducing conditions; stability fields for precipitation of goethite, magnetite, pyrite, and pyrrhotite are shown; normal seawater conditions are within the stippled region. Redrawn from Henshaw and Merrill (1980) with permission of the American Geophysical Union.

However, diagenetic alteration of detrital ferromagnetic minerals can take place in the upper few meters of hemipelagic sediments (Karlin and Levi, 1985). If a high sedimentation rate prevents complete oxidation of organic matter prior to burial, a two-layer system develops with an oxidizing upper layer less than 1 m thick overlying anoxic sediment below. Figure 8.6 suggests that these reducing conditions could drive the Fe–S–H_2O system into the pyrite stability field. Indeed, the magnetite content of organic-rich hemipelagic muds has been observed to decrease by at least a factor of 10 in the upper meter (Figure 8.7). This decrease in magnetite content and attendant NRM are caused by dissolution of detrital magnetite with accompanying precipitation of pyrite. If this *sulfurization* completely dissolves the detrital magnetite, the original DRM is destroyed.

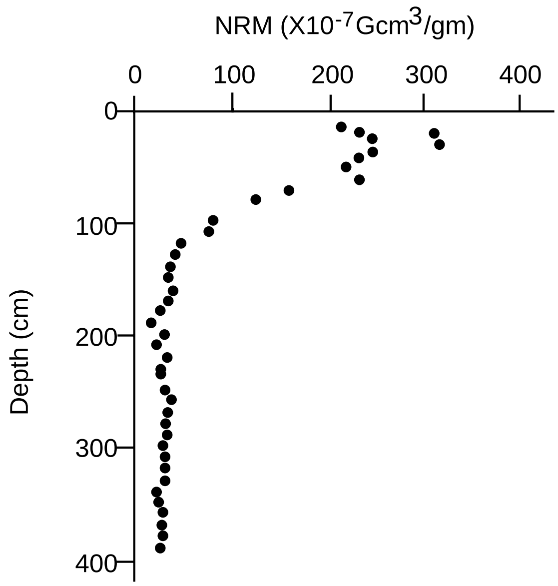

Figure 8.7 NRM intensity versus depth in a core of hemipelagic marine sediment. The core was collected from the lower continental shelf off the coast of Oregon in 1820-m water depth; the sediment is olive green, heavily bioturbated, suboxic hemipelagic mud; the mean sediment accumulation rate was ~120 cm/1000 yr; NRM intensity is after alternating-field demagnetization to peak field of 150 Oe (15 mT). Redrawn from Karlin and Levi (1985), with permission of the American Geophysical Union.

Fortunately, a significant fraction of the detrital magnetite usually survives until anoxic reactions decrease or are halted by cementation or lithification. In strongly reducing environments, however, detrital magnetite may be totally destroyed or survive only within early-formed concretions. Marine sediments with high sulfide content thus are unattractive targets for paleomagnetic study.

Pelagic sediments

Over half the ocean floor is covered by pelagic sediments that are primarily calcareous, diatomaceous, or radiolarian oozes. Gradual lithification and cementation take place by dissolution and recrystallization of foraminifera and coccoliths. Rates of sediment accumulation for pelagic sediments are only a few mm/1000 yr, and conditions are more uniformly oxidizing than for hemipelagic sediments. Detrital magnetite and titanomagnetite constitute about 0.01% by volume.

Fossil-bearing pelagic sediments are commonly reliable paleomagnetic recorders, whereas pelagic sediments without recognizable fossils tend to yield paleomagnetic records that progressively deteriorate in quality down the core (Henshaw and Merrill, 1980). Two diagenetic processes are thought to be responsible:

1. Progressive low-temperature oxidation of detrital magnetite often yields maghemite. This process might be particularly important for pelagic red clays common in the North Pacific. Organic matter in fossil-bearing pelagic sediments might prevent oxidation and account for the superior quality of paleomagnetic records from fossil-bearing sediments.
2. Authigenic precipitation of ferromagnetic ferromanganese oxides produce a slowly acquired CRM that overprints the original DRM.

Ancient Limestones

A detailed review of rock magnetism and paleomagnetism of marine limestones is given by Lowrie and Heller (1982). Only the basic properties are described here.

Some limestones are paleomagnetic recorders of extraordinary fidelity, while others yield little useful paleomagnetic information. Common ferromagnetic minerals in marine limestones included magnetite, goethite, hematite, and maghemite. With the exception of limestones suffering wholesale chemical remagnetization during orogenesis, morphology and chemistry of grains indicate that the magnetite is detrital. The primary paleomagnetism in most limestones is a pDRM carried by detrital magnetite.

Hematite is present as a pigment in red and pink limestones. Some detailed examinations have shown that hematite pigment can form as an early diagenetic product from goethite. In such rocks, CRM carried by the hematite can be essentially contemporaneous with DRM carried by detrital magnetite. However, if significant hematite is present, relative timing of DRM carried by magnetite and CRM carried by hematite must be established on an individual case basis.

Goethite is widespread in limestones and coexists with both magnetite and hematite. The presence of significant goethite is usually ominous for paleomagnetic investigations. Goethite can precipitate directly from solution (Figure 8.6) or result from alteration of pyrite, which is particularly common in white and blue-gray limestone. This alteration may be diagenetic but can also occur during subaerial weathering of porous limestone. Goethite often carries an unstable magnetization and dehydrates to hematite during laboratory heating to 300°C, leading to major complications during thermal demagnetization experiments. Thus, the presence of significant goethite generally leads to difficulties in isolating primary DRM carried by magnetite.

For many limestones, laboratory heating to 450° to 650°C produces new magnetite, either from pyrite or by reduction of hematite. This magnetite has superparamagnetic grain size and rapidly acquires troublesome VRM components that complicate isolation of primary DRM. Limestones with significant detrital magnetite but without significant pyrite or goethite can yield highly reliable paleomagnetic data. However, presence of significant pyrite or goethite usually leads to insurmountable difficulties. The most advantageous sedimentary environment for retaining primary DRM in pelagic limestones is a slightly oxidizing environment in which rapid cementation halts diagenetic changes, preserving detrital magnetite and preventing production of goethite.

Laboratory evidence that the remanent magnetization of a limestone is carried by magnetite is a necessary but not sufficient condition to assert that the magnetization is a primary DRM. As discussed below, secondary authigenic magnetite has been found in some Paleozoic limestones. Especially for ancient limestones that have been subjected to complex geochemical and tectonic history, field tests of paleomagnetic stability are indispensable.

MAGNETIC ANISOTROPY

Rocks in which intensity of magnetization (whether induced or remanent) depends on direction of the applied magnetic field have *magnetic anisotropy*. In such rocks, the direction of magnetization can deviate from that of the magnetizing field. There are two kinds of magnetic anisotropy:

1. *anisotropy of magnetic susceptibility* (AMS), in which susceptibility is a function of direction of the applied field; and
2. *anisotropy of remanent magnetization*, in which acquired remanent magnetization may deviate from the direction of the magnetic field at the time of remanence acquisition. Anisotropy of remanent magnetization has obvious implications for the accuracy of paleomagnetic records.

Studies of anisotropy of magnetic susceptibility have a wide range of applications (Hrouda, 1982; MacDonald and Ellwood, 1987). AMS exceeding 5% is generally observed only in rocks with obvious megascopic fabric, and values exceeding 10% are rare. But AMS of a few percent can be easily measured.

Because AMS can be measured more quickly and easily than, for example, measuring mineral orientations by optical analysis of thin sections, AMS has been used to examine development of petrofabrics.

Anisotropy of magnetic susceptibility is commonly expressed by comparing magnetic susceptibility values in three mutually perpendicular directions: K_1 = maximum susceptibility; K_2 = intermediate susceptibility; K_3 = minimum susceptibility. These values describe the *magnetic susceptibility ellipsoid*. If $K_1 = K_2 = K_3$, the ellipsoid is spherical; if $K_1 \approx K_2$ but $K_2 > K_3$, the ellipsoid is oblate (flattened); if $K_1 > K_2$ and $K_2 \approx K_3$, the ellipsoid is prolate (cigar-shaped). Magnetic susceptibility ellipsoids are usually interpreted as indicating statistical alignment of elongate or platy magnetic grains, usually ferromagnetic grains. For example, elongate magnetite grains in a rock with a pronounced foliation will have long axes rotated toward the foliation plane. The resulting magnetic susceptibility ellipsoid is oblate with K_3 perpendicular to foliation. Conversely, a rock with significant lineation will have a prolate magnetic susceptibility ellipsoid with K_1 parallel to the lineation direction.

AMS applications have been made to sedimentology, igneous processes, and structural geology. Sedimentary rocks generally display a slight AMS of oblate susceptibility ellipsoid with K_3 perpendicular to bedding. AMS of sedimentary rocks can sometimes be used to determine paleocurrent directions (Ellwood, 1980; Flood et al., 1985). AMS has also proved useful in analyses of flow of volcanic rocks. Oblate magnetic susceptibility ellipsoids are often observed in volcanic rocks with flow fabrics; K_3 is found perpendicular to flow surfaces. Prolate magnetic susceptibility ellipsoids are sometimes observed with K_1 parallel to the lines of flow of volcanic rocks. In fact, AMS analyses can be used to locate source areas of volcanic rocks, especially ignimbrites and welded tuffs, by using the direction of the K_1 axis at widely separated sampling locations to triangulate on the source vent (Ellwood, 1982; Knight et al., 1986).

In structural applications, AMS has been used to examine patterns of strain. An oversimplified view is that elongate ferromagnetic grains are passively rotated during straining of rocks. For example, the pattern of AMS in a shear zone might be used to decipher the strain involved. Applications to mylonite zones have been reported by Goldstein and Brown (1988) and Ruf et al. (1988). Quantitative relationships between strain and AMS are needed to infer strain directly from AMS. Kligfield et al. (1983) have developed such a relationship for Permian red sediments of the Maritime Alps.

Rocks with substantial AMS are likely to be anisotropic for acquisition of remanent magnetism and therefore not accurate paleomagnetic recorders. Many rocks that are of interest for AMS studies have obvious petrofabrics, which indicate that they are not appropriate for paleomagnetic analysis. But how much AMS can be tolerated? A useful generality is that paleomagnetic data from rocks with AMS exceeding about 5% should be viewed with particular caution. However, in the case of magnetite-bearing rocks, AMS is dominated by multidomain grains while single-domain and pseudosingle-domain grains are the paleomagnetic recorders. So AMS might not be closely related to anisotropy of remanent magnetization (Stephenson et al., 1986).

Because conditions of primary NRM formation are indirectly inferred and difficult to reproduce, anisotropy of remanent magnetization must be examined indirectly. Some volcanic rocks with pervasive flow fabric have significant deflection of TRM from the direction of the magnetic field present during cooling. However, these cases are rare, and significant anisotropy of remanent magnetization in the vast majority of igneous rocks or in red sediments is demonstrably absent or unlikely.

Most recent attention has focused on sedimentary rocks, especially those with possible inclination error. Some interesting observations have been made by using a form of remanent magnetization that can be easily produced in the laboratory. *Anhysteretic remanent magnetization* (ARM) is produced by superimposing an alternating magnetic field (e.g., Figure 5.1a) on a small direct magnetic field. The ferromagnetic grains that carry ARM are those grains with microscopic coercive force up to the maximum amplitude of the alternating magnetic field used to impart the ARM. As with other forms of remanent magnetization, SD and PSD grains are more effective carriers of ARM than are MD grains. So imparting ARM in different directions within a rock sample allows examination of fabric in the important carriers of remanent magnetism, the SD and PSD grains.

Observed anisotropy of ARM (more or less ARM acquired in some sample directions than in other directions) indicates possible anisotropy in acquiring NRM. This provides a warning that the rock might not be an accurate paleomagnetic recorder. Also ARM can be measured for weakly magnetic rocks (such as limestones), whereas AMS can be measured only for rocks with substantial ferromagnetic content (McCabe et al., 1985; Jackson et al., 1988). Potential applications in deciphering possible inclination error in sedimentary rocks are of major significance.

CHEMICAL REMAGNETIZATION

To this point, secondary CRM components have been discussed only in the section on magnetization of marine sediments. However, many rocks suffer chemical remagnetization in which primary NRM is destroyed and replaced by secondary CRM. In this section, some examples of remagnetization are discussed. This is definitely a "good news and bad news" situation. The bad news is that remagnetized rocks do not retain a primary NRM and many objectives of paleomagnetic study of these rocks cannot be met. The good news is that the timing and processes of remagnetization are providing important insights into orogenic and geochemical processes.

Weathering can affect original ferromagnetic minerals and result in the formation of new ferromagnetic minerals with attendant CRM components. Because surface conditions are predominantly oxidizing, reactions that transform primary ferromagnetic minerals (such as magnetite) to higher oxidation state minerals (such as hematite or goethite) are common. Although the usual concern is for CRM acquired during recent weathering, secondary CRM components may have resulted from ancient weathering. A clear case of remagnetization of older rocks by ancient weathering was presented by Schmidt and Embleton (1976).

Regional lateritization of western Australia in Late Oligocene to Early Miocene time produced chemical remagnetization of Late Paleozoic through Mesozoic strata. Lateritization and acquisition of CRM in resulting hematite occurred over a time interval spanning at least one geomagnetic polarity reversal because both normal- and reversed-polarity CRM is observed. The paleomagnetic pole determined from the direction of chemical remagnetization coincides with the 20 to 25 Ma pole position for Australia. This inferred age of chemical remagnetization in western Australia is supported by independent paleoclimatological and geochronological data indicating a Late Oligocene to Early Miocene interval of peneplanation and lateritization in northern and western Australia.

The most intensely studied remagnetization is that of Early and Middle Paleozoic rocks in the Appalachian region of eastern North America. This remagnetization took place during the Late Carboniferous and Permian, affected a wide variety of rock types, and is clearly related to the Late Paleozoic Alleghenian Orogeny. An excellent review article was provided by McCabe and Elmore (1989).

Creer (1968) observed that many paleomagnetic poles from Early Paleozoic rocks of North America were similar to poles from Late Paleozoic rocks. He suggested that the Early Paleozoic units were chemically remagnetized in the Late Paleozoic by protracted weathering while North America was situated in tropical paleolatitudes (see Chapter 10). As more paleomagnetic data were obtained and more sophisticated demagnetization techniques and analyses were applied, multiple components of NRM were observed in Early Paleozoic units of the Appalachians. For example, Van der Voo and French (1977) found two components of NRM in the Ordovician Juniata Formation. The highest-stability component passed a fold test and is therefore prefolding. But a lower-stability component was found to fail the fold test, with in situ directions indicating a Late Paleozoic age. Van der Voo and French (1977) argued that this Late Paleozoic component of NRM was the result of remagnetization by thermal and/or chemical effects associated with the Alleghenian Orogeny rather than the result of surface weathering.

Subsequent studies have documented the widespread nature of this remagnetization. Irving and Strong (1984, 1985) observed both prefolding and postfolding components of NRM in Early Carboniferous red sediments of western Newfoundland. This observation led to significant revision of ideas about tectonic motions of terranes in the Appalachians, and many of the remagnetizations have been shown to be synfolding (Chapter 5, Figure 5.13), indicating a causal connection with the Alleghenian Orogeny.

Before detailed analysis of remagnetized limestones in the Appalachians, it was commonly believed that only oxidation reactions could lead to remagnetization. But Late Paleozoic remagnetizations of some Appalachian limestones are carried by authigenic magnetite (Scotese et al., 1982; McCabe et al., 1983). Magnetite has been separated from the remagnetized limestones and identified as authigenic by (1) lack of Ti or other Fe-substituting cations that are commonly found in magnetite from igneous or extraterrestrial sources and (2) hollow or botryoidal morphology indicating in situ precipitation (Figure 8.5b). Independent evidence indicates that precipitation occurred at low temperature (<200°C). Recent observations have revealed magnetite crystals with pyrite cores, indicating that authigenic magnetite is an alteration product of preexisting pyrite (Suk et al., 1990).

The geochemistry of this remagnetization is complex; remagnetizations in red sediments are carried by hematite, whereas remagnetizations in most carbonates are carried by magnetite. Furthermore, not all researchers agree that remagnetization is necessarily the result of chemical reactions leading to CRM. Kent (1985) concluded that thermoviscous effects of burial are important (remember TVRM from Chapter 3?). Also van der Pluijm (1987) and Kodama (1988) argue that strain effects during folding play an important role in altering the NRM of some units.

The role of fluids in producing the chemical remagnetizations is also of interest. Lateral migration of “orogenic fluids” may result from motions of thrust sheets driving fluids toward the craton (McCabe et al., 1983; Oliver, 1986). In favorable circumstances, the directions of the remagnetization can be used to date the time of fluid migration and orogeny. This possibility is given economic incentive because the fluids involved include hydrocarbons. Authigenic magnetite has been found in bitumen of remagnetized Paleozoic carbonates in the midcontinent region of North America (McCabe et al., 1987). Evidence for Cretaceous remagnetization carried by authigenic magnetite associated with hydrocarbon migration has been found in the Rocky Mountain region (Benthien and Elmore, 1987). The possible use of remagnetizations carried by authigenic magnetite as a technique for dating hydrocarbon migration is under investigation.

There are several lessons to be gained from this discussion of chemical remagnetization:

1. Detailed demagnetization analyses are essential to resolve the multiple components of NRM that are often encountered in old rocks that have experienced complex histories.
2. Field tests of paleomagnetic stability can provide crucial information about acquisition times for these components of NRM.
3. Geochemical and thermal effects of orogeny can lead to remagnetization by a variety of mechanisms.
4. Rock-magnetic and paleomagnetic analysis of the remagnetization process can lead to new applications of the paleomagnetic technique.

Discovery of secondary CRM is rarely the intent of a paleomagnetic study. But the direction of chemical remagnetization can constitute an important observation potentially allowing determination of the age of geochemical events such as orogenic fluid motions or hydrocarbon migration. As noted by McCabe and Elmore (1989), “Paleomagnetic studies promise to be important in assessing the role of orogeny in driving fluid migrations within sedimentary basins and in constraining the age of the migrations and the nature of the fluids.”

THE RED BED CONTROVERSY

Intensity of natural remanent magnetization in red sediments is commonly $\geq 10^{-5}$ G (10^{-2} A/m). These intensities can be measured on a variety of instruments that are available from the early development of paleomagnetism, and red sediments are abundant in the stratigraphic records of most continents. Accordingly, numerous paleomagnetic studies have been undertaken on red sediments. However, there are major differences of interpretation about magnetization acquisition by red sediments. The resulting debate is the "red bed controversy."

The extreme views can be summarized as follows:

1. High-stability components of NRM (the ChRM) in red sediments are carried by detrital specular hematite that is magnetized by DRM processes (Elston and Purucker, 1979; Steiner, 1983). This ChRM is penecontemporaneous with deposition and can provide high-fidelity records of the paleomagnetic field, including records of paleosecular variation (Baag and Helsley, 1974) and geomagnetic polarity transitions (Herrero-Bervera and Helsley, 1983; Shive et al., 1984).
2. Multiple components of CRM are acquired during protracted chemical processes occurring up to 10 m.y. after deposition of a red sediment (Roy and Park, 1972; Larson and Walker, 1975; Turner, 1980; Walker et al., 1981; Larson et al., 1982). Neither useful polarity stratigraphy nor records of paleosecular variation or geomagnetic polarity transitions can be retrieved from red sediments.

The fundamental question can be stated as follows: "Is the ChRM in red sediments a DRM acquired penecontemporaneously with deposition, or is it a CRM acquired during protracted chemical change occurring up to 10 m.y. post-deposition?" The answer is a resounding, authoritative "Yes!" Obviously, there would be no controversy if the situation were simple. The discussion below does not provide an answer to the red bed controversy but rather explains the fundamental evidences and arguments. On each aspect of the controversy, the discussion proceeds from generally accepted background information to more controversial interpretations.

The dominant (usually exclusive) ferromagnetic mineral in red sediments is hematite, which occurs in two general categories:

1. Opaque crystals that are generally larger than 1 μm and exhibit silvery, anisotropic reflectance when viewed in polished section. This form is specular hematite, or simply specularite (Figures 8.8a, 8.8b, 8.8c).
2. Fine-grained (< 1 μm) hematite pigment that is translucent and is largely responsible for the red coloration of the sedimentary rock (Figure 8.8d).

Pigmentary hematite often cements detrital grains and is clearly formed by postdepositional chemical processes. Two important reactions probably account for a majority of hematite pigment: (1) dehydration of ferric oxyhydroxides in newly deposited sediment and in soil layers and (2) alteration of Fe-bearing silicates. Textural relationships sometimes allow the sequence of pigment-forming reactions to be determined. However, the rates of these reactions are not sufficiently known to allow secure statements about the time interval required for pigment formation.

Because of the small grain size of many of the pigment crystals, the magnetization of many pigment grains is unstable over geologic time, and these grains tend to acquire viscous magnetization. Usually, this VRM can be erased by either chemical or thermal demagnetization techniques. Pigmentary hematite also can acquire CRM during precipitation and grain growth in the geomagnetic field. But the difficulty in interpretation of this CRM is determining its time of formation. Some experiments indicate that CRM carried by hematite pigment is composed of multiple components of magnetization acquired during protracted chemical precipitation, perhaps millions of years after deposition (Roy and Park, 1972). Although certainly not a universal observation, several studies have shown that the ChRM is carried not by the hematite pigment but rather by the specular hematite (Collinson, 1974; Tauxe et al., 1980). So for most red sediments, the question of the timing of ChRM acquisition becomes a question of time of formation of the specular hematite.

The major question is whether the specular hematite in a red sediment was deposited as a detrital grain of specular hematite and could potentially have acquired a DRM or formed by postdepositional oxidation of magnetite or other Fe-bearing minerals and therefore carries a CRM that could have been acquired long

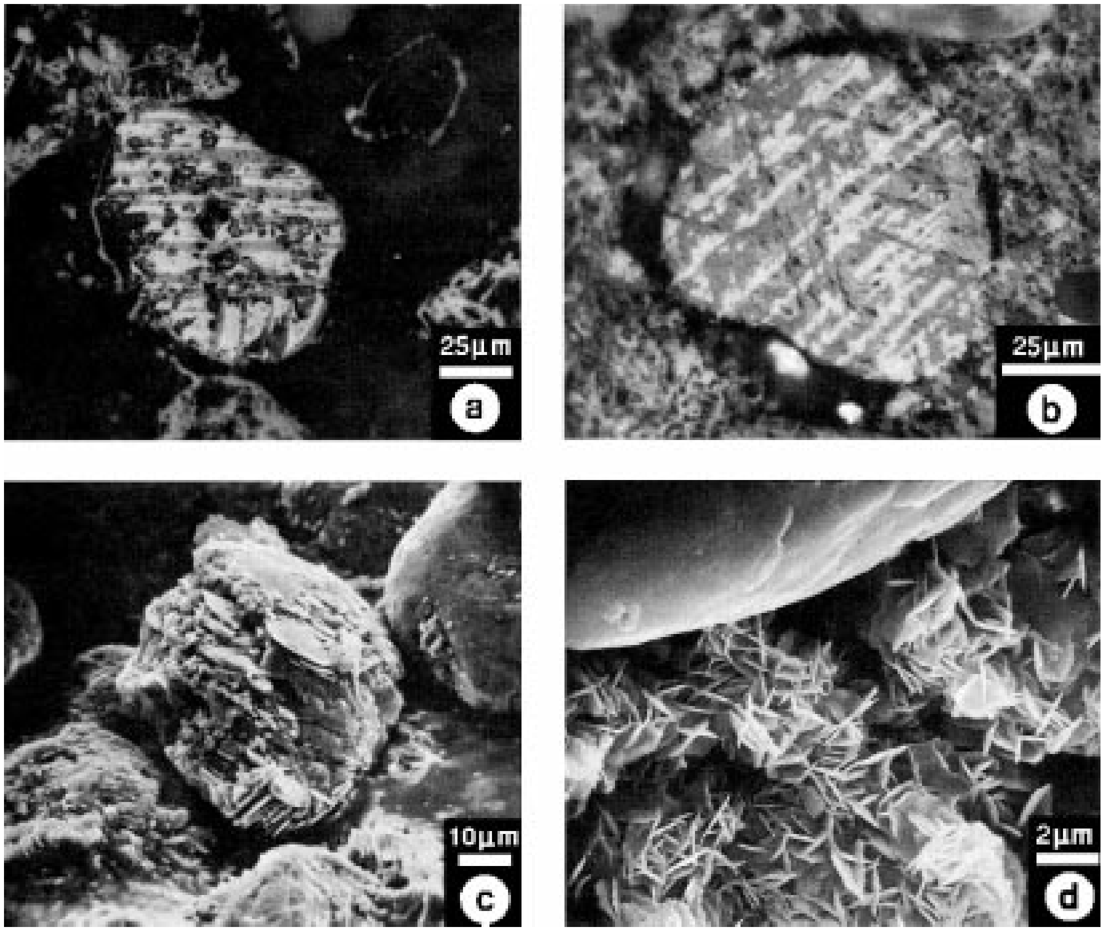

Figure 8.8 Optical and SEM photomicrographs of hematite and associated minerals in red sediments. (**a**) Detrital grain exhibiting "tiger-striped" ilmenite-hematite intergrowth; darker ilmenite layers alternate with lighter layers of hematite. (**b**) Polycrystalline martite grain showing crystalline units intersecting along octahedral planes inherited from replacement of parent magnetite; the entire grain is hematite; differing shades of gray result from different crystallographic directions for different portions of the grain; (**c**) SEM photomicrograph of martite grain with overgrowths of authigenic specular hematite. (**d**) SEM photomicrograph of interlocking hematite crystals within a sand-size void. All samples are from the Wupatki Member of the Moenkopi Formation. Photomicrographs kindly provided by T. Walker.

after deposition. To appreciate the difficulty of addressing this question, we must consider the possible origins of specular hematite in general.

The forms of specular hematite present in red sediments include the following:

1. *Igneous/metamorphic specular hematite.* As discussed in Chapter 2, hematite can result from igneous processes. Grains that are intergrowths of hematite and ilmenite resulting from high-temperature exsolution are occasionally found in red sediments. These grains often exhibit a "tiger-striped" texture (Figure 8.8a). Such intergrowth grains result from high-temperature processes and must have been eroded from an igneous source terrane and deposited as specular hematite.
2. *Martite.* Grains of specular hematite often show clear evidence of resulting from oxidation of preexisting magnetite. Pseudomorphs of the original magnetite are preserved, and these grains contain ilmenite laths resulting from deuteric oxidation of the original titanomagnetite grain (Figure 8.8b). Composite grains with specular hematite exteriors and magnetite cores are also observed. Grains of specular hematite with clear evidence of formation by oxidation of magnetite are referred to as

martite. But observation of martite grains does not necessarily provide evidence that the grain was martite when deposited; it could have been martitized by in situ postdepositional oxidation of a detrital magnetite grain. Sometimes delicate authigenic overgrowths of specular hematite exist on martite grains (Figure 8.8c). At least the overgrowth portions of these grains must have resulted from postdepositional authigenesis. However, the time of oxidation of most martite grains is indeterminate from petrographic analysis alone.

3. *Specular hematite in Fe-bearing silicates*. Oxidation of Fe-bearing silicates often yields specular hematite that may form in cleavage planes of the host mineral. Textural evidence often indicates sequences of reactions that are the result of in situ oxidation (Walker et al., 1981).
4. *Specular hematite of uncertain origin*. Many grains of specular hematite lack textural patterns that provide information about their origin.

Given the difficulty of determining the origin of specularite grains in red sediments, it is not surprising that disparate interpretations exist. Recent sedimentary deposits do not often contain specular hematite as the dominant ferromagnetic mineral; magnetite is usually dominant (Van Houten, 1968). This observation has been used to argue that most specular hematite must be formed by postdepositional oxidation of detrital magnetite. However, some modern streams do deposit detrital specularite, and these deposits do possess a substantial DRM (Tauxe and Kent, 1984). This DRM is further observed to have a pronounced inclination error, which probably results from the low ratio of magnetic moment to gravitational torque. Paleomagnetic studies of some ancient red sediments has revealed inclinations of magnetization that seem to be systematically shallowed in heavy mineral layers containing high concentrations of specular hematite (Elston and Purucker, 1979; Steiner, 1983). So some evidence favors postdepositional formation of the majority of specularite, while other evidence indicates the possibility of DRM or pDRM acquisition in detrital specularite.

The best evidence for the mode and timing of acquisition of NRM by red sediments comes from field tests (Chapter 5) applied to sedimentary structures. During deposition of stratigraphic sequences of red sediments, rip-up clasts of previously deposited layers are occasionally incorporated within intraformational conglomeratic layers. Oriented samples of these clasts can be used as a conglomerate test. If directions of ChRM in numerous clasts are randomly directed, the magnetization must have been acquired before the layer yielding the clasts was disrupted. This test has been applied to Mesozoic red beds of western North America with mixed results. Some conglomeratic layers pass the conglomerate test, and others appear to fail this test (Purucker et al., 1980; Liebes and Shive, 1982; Larson and Walker, 1982).

Soft-sediment deformational structures such as load casts and slump folds have also been investigated. For the Triassic Moenkopi and Chugwater formations of the Rocky Mountains and Colorado Plateau, magnetizations of most load casts and small-scale slump folds (<1 m amplitude) were found to fail the fold test, while larger-amplitude folds yielded magnetizations that passed the fold test (Liebes and Shive, 1982). These observations indicate that the ChRM of these Mesozoic red sedimentary formations was formed after deposition but prior to burial by about 1 m of sediment. The conclusion was that the ChRM is a CRM acquired predominantly within a few hundred years of deposition.

Examination of the within-site and between-site dispersion of ChRM directions can also provide information about the time interval over which this magnetization was acquired. Acquisition over a time interval exceeding 10^5 yr would yield site-mean ChRM directions with angular dispersion much lower than the dispersion expected for sampling geomagnetic secular variation. However, if dispersion of site-mean ChRM directions <u>between</u> stratigraphically superposed sites substantially exceeds dispersion <u>within</u> individual beds, some directional dispersion from sampling geomagnetic secular variation was probably recorded. Although detailed examinations are not numerous, observed between-site dispersion in some red sediments indicates acquisition of ChRM within 10^2 to 10^3 yr of deposition (Ekstrand and Butler, 1989). Herrero-Bervera and Helsley (1983) and Shive et al. (1984) investigated a polarity transition (~1-m-thick stratigraphic interval deposited while the geomagnetic field was switching polarity) within the Chugwater Formation. They found that detailed directional changes of ChRM were consistent in multiple stratigraphic sections over a distance of 1 km. They argued that these consistent observations of rapid directional changes of the geomagnetic field require that the ChRM was acquired within ~10^2 yr of deposition.

So the present situation is that some red sediments appear to contain ChRM that is well defined, without multiple components. The bulk of available evidence suggests that these red sediments acquire a ChRM within 10^2 yr of deposition, most likely as a CRM. However, other red sediments with more complex magnetizations show evidence of components of magnetization acquired long after deposition, although exactly how long is not well constrained.

Figure 8.9 presents a schematic view of magnetization processes in red sediments. The time scale on the ordinate is a “best guess” at the time intervals over which different mechanisms of magnetization may operate. There are two basic categories:

1. *Detrital remanent magnetization*. Depositional or postdepositional DRM could form if a significant portion of the specular hematite is detrital. Mineralogically mature sediment would be more likely to contain detrital specular hematite than would first-generation (mineralogically immature) sedimentary rocks being eroded form a nearby igneous source terrane. Although still a matter of debate, it is a minority view that DRM is the major origin of ChRM in red sediments.
2. *Chemical remanent magnetism*. CRM is acquired during martitization of detrital magnetite, formation of specular hematite from Fe-bearing silicates, and authigenic production of pigmentary hematite. We could divide the CRM field into two subregions: (a) “early” chemical remanence, referring to CRM formed within 10^2 to 10^5 yr of deposition, and (b) “prolonged” chemical remanence, referring to CRM formed over longer time intervals. This subdivision has some paleomagnetic utility because early CRM could be applied to magnetic polarity stratigraphy when acquisition of the characteristic NRM within 10^5 yr of deposition is important. However, CRM formed over intervals up to perhaps 10^7 yr could still be used to determine paleomagnetic poles.

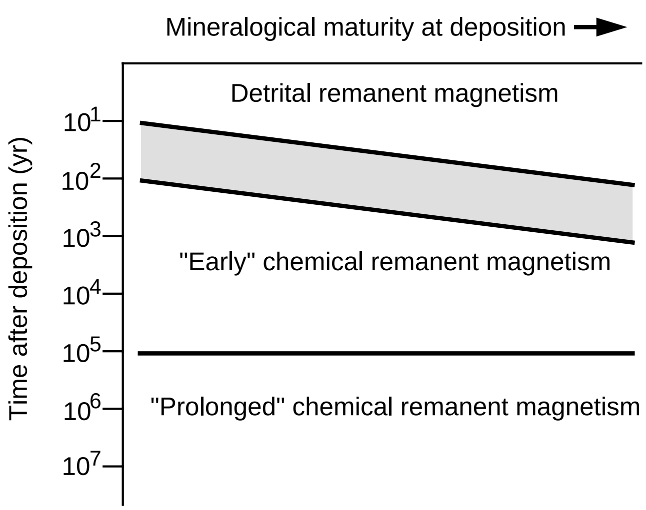

Figure 8.9 Mechanisms of magnetization in red sediments. Mineralogical maturity relates to the oxidation state of the deposited sediment; highly oxidized sediments have higher mineralogical maturity; demarcations between fields of different magnetization mechanisms are highly schematic. Adapted from Turner (1980).

While there are many uncertainties about the magnetization processes in red sediments, there are a number of factors that certainly play a role:

1. *Mineralogical maturity at deposition*. Immature sediments with abundant low-oxidation-state minerals might experience rapid oxidation and acquire the majority of their CRM quickly. More mature sediments might require more time for these postdepositional chemical reactions. This tendency for mineralogically immature sediments to more quickly acquire CRM is indicated schematically in Figure 8.9 by the sloping interface between the DRM and “early” CRM fields.

2. *Grain size of sediment*. Finer-grained sediments have particles with larger surface-to-volume ratio that likely undergo chemical changes more rapidly than larger-grained sediments. Clay diagenesis and cementation occur more rapidly in finer-grained sediments. So CRM in fine-grained, cemented red sediments might be acquired more quickly than in coarse-grained sediments.
3. *Depositional environment and paleoclimate*. A depositional environment that is highly oxygenated will produce more rapid oxidation favoring early formation of CRM. Warm, moist conditions yield more rapid and continual CRM formation than arid, dry conditions.

The bottom line on this discussion of magnetization of red sediments is that the processes are indeed complex and still controversial. Paleomagnetic data obtained from red sediments must be evaluated on a case-by-case basis. The best evidence for timing of remanence acquisition comes from field tests of paleomagnetic stability. The consensus view is that red sediments with uncomplicated, high-stability ChRM probably acquired this magnetization by CRM processes that occurred within 10^3 yr after deposition. Paleomagnetic data from these red sediments are useful for magnetic polarity stratigraphy and for determination of paleomagnetic poles. The timing of magnetization components for red sediments that yield complex, multicomponent NRM is poorly constrained; caution must be exercised in interpreting such results.

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