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Cross-references

Demagnetization
 Magnetic Susceptibility, Anisotropy
 Magnetization, Isothermal Remanent (IRM)
 Magnetization, Thermoremanent (TRM)

MAGNETIZATION, CHEMICAL REMANENT (CRM)

Chemical remanent magnetism (CRM) is imparted to ferro- and ferri-magnetic minerals by chemical processes, at temperatures below their Curie points, in the presence of an effective magnetic field. Here, chemical processes are considered broadly to include but not be limited to modifications in oxidation state, phase changes and crystal growth. The effective magnetic field is the resultant vector field acting on the chemically-altered material, including the external and various interaction fields.

Nearly a century and half ago, Beetz (1860) discovered CRM during laboratory electrolytic depositions of iron; these observations supported Weber's hypothesis that some atoms possess intrinsic magnetization. These results were confirmed by Maurain (1901 and 1902) with electrolytic depositions of iron and nickel in the presence of external fields. Koenigsberger (1938, part 1, p. 122 & part 2, p. 319) noted the presence of coherent remanence in some sedimentary rocks, which he called *crystallization remanence*, and he advanced the hypothesis that it was "impressed by the Earth's field at temperatures between about 100°C and 500°C during the time of lattice changes in magnetite which result very probably from unmixing of Fe₂O₃."

With the rapid growth of paleomagnetism after World War II, it became apparent to many rock magnetists that the remanence of many sediments, especially red beds, was at least partly controlled by magnetic minerals chemically precipitated subsequent to deposition (e.g., Blakett, 1956).

In rocks, CRM is usually a secondary remanence; this is an important reason that in geophysics CRM studies lag behind investigations of TRM (Thermoremanent magnetization) and DRM (Depositional remanence), which are usually responsible for the primary remanence. In most paleomagnetic studies, it is advantageous to select samples that retain their primary TRM or DRM; secondary CRM is a nuisance to be avoided, if possible. However, in nature, CRM can rarely be entirely neglected. If a secondary CRM is superimposed on an extant primary TRM or DRM, one of the first objectives would be to remove (demagnetize) the CRM in order to expose the primary remanence. In many cases, especially for older, chemically-altered sediments, CRM is the dominant characteristic remanence, the primary DRM having been obliterated by processes similar to those that are responsible for the CRM.

Theory for CRM

Many CRM properties can be explained with Néel's (1949, 1955) thermal fluctuations theory for super-paramagnetic (SP) particles, which has been highly successful at explaining TRM and many other properties of remanence. We consider the an assemblage of non-interacting, uniformly magnetized SP particles with identical volumes, v , which have uniaxial anisotropy and aligned easy axes. For thermal equilibrium and in the presence of an external magnetic field, H , applied parallel to the easy axes, the volume magnetization, M , is given by the equation

$$M = nM_S(T) \tanh[M_S(T)vH/kT] \quad (\text{Eq. 1})$$

The argument $[M_S(T)vH/kT]$ is the alignment factor, which vanishes for $H = 0$. $M_S(T)$ is the spontaneous saturation magnetization of the grains at temperature T ; n is the number of magnetic particles per unit volume; k is Boltzmann's constant; T , the absolute temperature ($^{\circ}\text{K}$).

When the field is removed, M decays, seeking the new state of thermal equilibrium, $M = 0$, at a rate determined by the relaxation time, τ . M is given by the equation

$$M = M_0 \exp[-t/\tau] \quad (\text{Eq. 2})$$

where M_0 represents the initial magnetization; t is the elapsed time since the field was removed. For the particle assemblage described above, Néel derived an equation for τ , and in zero external magnetic field,

$$\tau = C^{-1} \exp[Kv/kT] \quad (\text{Eq. 3})$$

where K is the uniaxial anisotropy energy per unit volume; C is the frequency factor whose value is on the order of 10^9 s^{-1} . C is a function of T and T -dependent material properties; however, C 's variation with temperature is significantly less than the exponential factor, and, in comparison, C is usually treated as constant. By contrast, τ , which is also a measure of the remanence stability, varies orders of magnitude in response to modest changes of the argument (Kv/kT).

We now consider isothermal CRM production at temperature T_A , caused by crystal growth of magnetic particles. The precipitating ferromagnetic particles grow from atomic/molecular paramagnetic nuclei to larger exchange-coupled SP grains with spontaneous magnetization, M_S . In an external field and thermal equilibrium, the particles are aligned according to Eq. (1). When the field is removed, and as long as the thermal fluctuations (kT) can easily overcome the anisotropy energy barriers (Kv), $\tau \ll \tau_L$, M decays quickly following Eq. (2), and there is no remanence. (τ_L is a characteristic laboratory time on the order of minutes.) As the particle volumes increase, it

becomes more difficult for thermal fluctuations to overcome the growing barriers to domain rotations. Because of the exponential dependence of τ on v in Eq. (3), the magnetization changes to a stable CRM over a very narrow range of volumes, Δv . At volume v_{AB} , $\tau \gg \tau_L$, and CRM is said to be blocked. v_{AB} is the critical blocking volume at T_A . That is, at temperature T_A , for $v < v_{AB}$, the particle assemblage is super-paramagnetic, while for $v > v_{AB}$, the magnetization is blocked as stable CRM in single domain (SD) particles. This is analogous to TRM production, where v is considered constant, and τ increases as T cools below the Curie point, T_C . Above the blocking temperature, T_B , the SP magnetic moments are aligned according to Eq. (1) with $\tau \ll \tau_L$, so that when the external field is removed, the magnetization quickly decays following Eq. (2). At lower temperatures $T < T_B$, $\tau \gg \tau_L$, and the magnetization is blocked as stable TRM.

CRM is usually difficult to distinguish from other remanences such as TRM and VRM (viscous remanence), because of overlapping stabilities and possible associations between CRM, partial-TRM (PTRM) and high-temperature VRM. For example, grain growth CRM at $T_A > T_R$, where T_R denotes room temperature, is in particles that have grown beyond the blocking volume, v_{AB} ; chemically precipitated particles with volumes $v < v_{AB}$ do not contribute to the CRM. On cooling from T_A to T_R , a PTRM will be produced in particles with blocking temperatures $T_A > T_B > T_R$. Also, ubiquitous time effects might contribute significant VRM superimposed on the CRM. In addition, the resultant remanence is likely to grow on cooling from T_A to T_R , due to the increase in M_S on cooling for most magnetic minerals. Such complexities make it difficult to uniquely isolate CRM from other remanences, hence it is probable that CRM occurrences in the paleomagnetic record are more numerous than is usually recognized.

As long as the magnetic ensemble consists of non-interacting, homogeneously magnetized SD particles, τ and the magnetic stability increase exponentially with particle volume. For multi-domain (MD) particles, τ decreases with increasing volume, principally because it becomes progressively easier to alter the remanence by domain wall movements, as opposed to rotating the magnetic moments of SD particles. These considerations of remanence stability and relaxation times were demonstrated experimentally by coercivity measurements of dispersed fine particles of ferromagnetic metals (Fe, Ni, Co), as the grain sizes increased due to progressive heat treatments (e.g., Meiklejohn, 1953; Becker, 1957). Zero initial coercivity in the SP region is followed by increasing coercivity with particle sizes, presumably in the SD size range; the coercivity then decreases for larger presumably MD particles (Figure M104). Haigh (1958) first applied Néel's theory and the grain size dependence of the coercivity to rock magnetism to explain CRM properties in growing particles of magnetite obtained during laboratory reduction of hematite. Kobayashi (1959) also examined CRM in magnetite obtained from hematite reduction and showed that CRM stability with respect to both alternating fields and thermal demagnetization was much greater than for IRM (isothermal remanence) and very similar to the stability of TRM (Figure M105). Kobayashi (1961) produced CRM in cobalt grains precipitated from Cu-Co alloy and showed that the specific CRM intensity had a similar bell-shaped grain size dependence as the coercivity, increasing from zero for SP particles, attaining a maximum value and then decreasing for inhomogeneously magnetized MD particles (Figure M106). Grain growth CRM can explain many examples in paleomagnetism, where the natural remanence (NRM) is predominantly a secondary remanence in chemically-altered rock and where new particles of a chemically-nucleated magnetic phase have grown beyond the SP size range.

CRM in Igneous Rocks

General Considerations

In igneous rocks, CRM can be produced by phase transformations or nucleation and growth of new magnetic minerals at temperatures

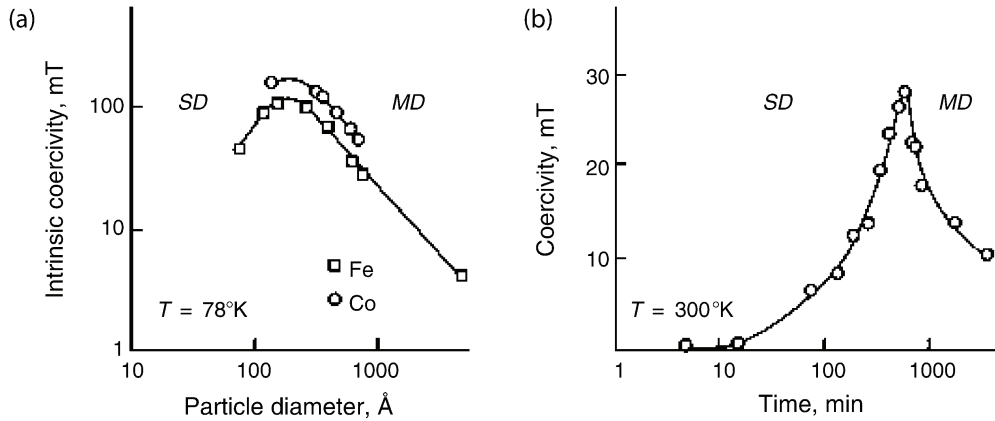


Figure M104 (a) Intrinsic coercive force of iron and cobalt as a function of particle size at liquid nitrogen temperature (modified from Meiklejohn, 1953). (b) Change in coercive force of Cu(98%)Co(2%) alloy as a function of annealing time (particle size) at 700°C, measured at 300 K (modified from Becker, 1957).

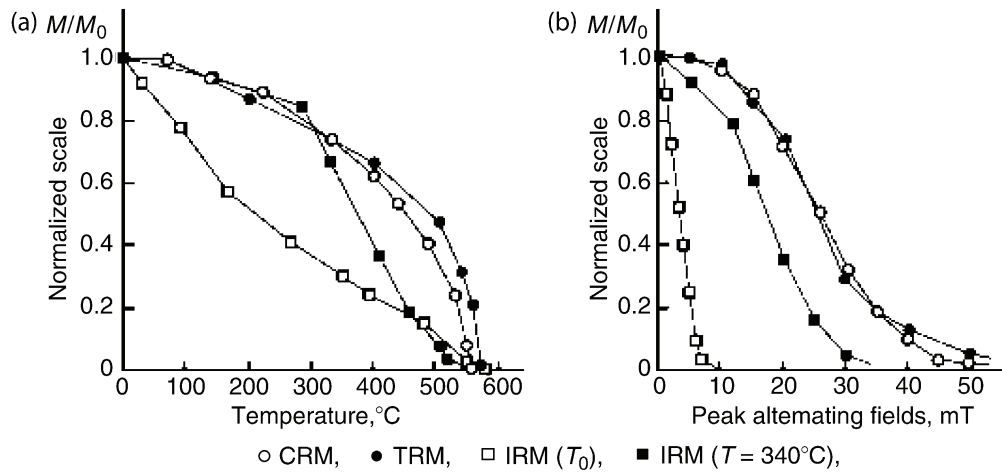


Figure M105 (a) Thermal, (b) alternating fields (AF) demagnetization of remanence in magnetite. All remanence measurements were made at room temperature. CRM was produced at 340°C with external fields of (a) 0.3 mT, (b) 1 mT. Total TRM was produced in (a) 0.3 mT, (b) 0.05 mT. IRM(T_0) at room temperature was produced in fields of (a) 20 mT, (b) 3 mT. IRM(T) was produced at 340°C in fields of (a) 2 mT, (b) 1 mT. CRM ($T = 340^\circ\text{C}$) and IRM ($T = 340^\circ\text{C}$) were cooled to T_0 in zero field prior to being demagnetized (modified from Kobayashi, 1959).

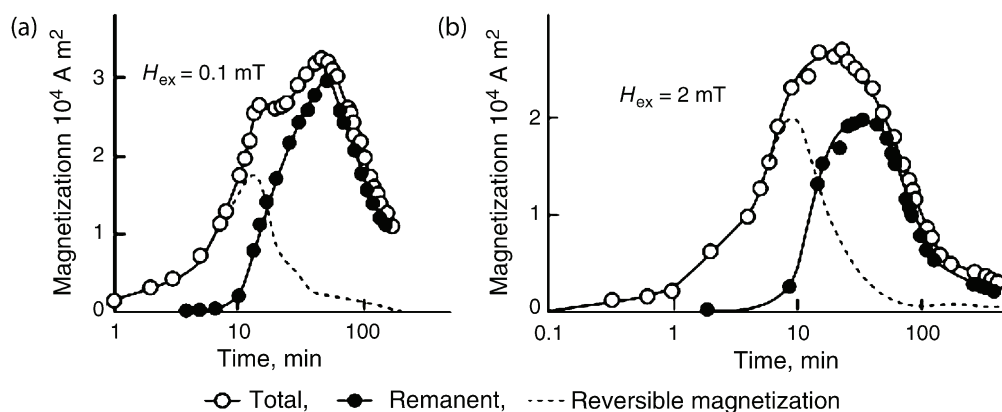


Figure M106 Total, remanent, and reversible magnetization of Cu-Co alloy as a function of annealing time (particle size) at 750°C, measured at 750°C. The reversible magnetization is the difference between the total magnetization (in presence of an applied field) and the remanence (modified from Kobayashi, 1961).

below the Curie point of the new magnetic species. The effective magnetic field at the site of the new magnetic particles determines the direction and intensity of the CRM. In igneous rocks, magnetic interactions among multiple phases, such as during exsolution of the iron-titanium solid solution series, might significantly modify the external field through magnetostatic or exchange interactions with phase(s) having higher blocking temperatures. Occasionally, such interactions might produce CRM with oblique directions or opposite polarity to the external magnetic field and the existing magnetic phase (negative magnetic interactions), and in rare cases a self-reversal may result (Néel, 1955).

CRM Origin of Marine Magnetic Anomalies

The first remanence of freshly extruded submarine basalts is TRM in stoichiometric titanomagnetites, $x\text{Fe}_2\text{TiO}_4(1-x)\text{Fe}_3\text{O}_4$, with $x \sim 0.6 \pm 0.1$, that is, about 60% molar ulvöspinel, with Curie points between $100^\circ\text{--}150^\circ\text{C}$ (Readman and O'Reilly, 1972). The initial magnetic phase is transformed at the sea floor by topotactic low temperature oxidation to cation deficient titanomaghemites. The oxidation is thought to proceed by a net cation migration out of the crystal lattice to accommodate a higher $\text{Fe}^{3+}/\text{Fe}^{2+}$ ratio and accompanying changing proportions of Fe:Ti. The resulting cation deficient phases retain the cubic crystal structure with smaller lattice dimensions, higher Curie points, approaching 500°C , and lower saturation magnetizations, responsible for the rapid diminution of the amplitudes of the marine magnetic anomalies away from spreading centers (e.g., Klitgord, 1976). Therefore, marine magnetic anomalies over the world's oceans can be considered to be preserved predominantly as CRM in oxidized Fe-Ti oxides.

The secondary CRM in submarine basalts usually retains the same polarity as the initial TRM recorded upon extrusion. This is indicated by the agreement, for overlapping time intervals, between the polarity time scales for marine magnetic anomalies, continental lavas and marine sediments, as well as the magnetic polarity of oriented dredged and drilled submarine basalts. The agreement of CRM and TRM directions is further supported by laboratory low temperature oxidation experiments of predominantly SD titanomagnetites (e.g., Marshall and Cox, 1971; Johnson and Merrill, 1974; Özdemir and Dunlop, 1985). The superexchange interactions vary with changes in cation distribution and lattice dimensions, as indicated by the reduced saturation magnetization and higher Curie points for the more oxidized titanomaghemites. However, it is possible that the orientations of the sub-lattice magnetic moments remain intact during low temperature oxidation of the titanomagnetite minerals, so that the ensuing CRM retains or inherits the original TRM direction.

Evidence from magnetic anomalies and magnetic properties of drilled oceanic basalts suggests an increase of the magnetization of extrusive submarine basalts of oceanic crust older than about 40 Ma (e.g., Johnson and Pariso, 1993). At present, the data are too sparse for one to be confident of the generality of this phenomenon or to select from several mechanisms that might be responsible.

CRM Influence on Paleointensity Studies

At present, only TRM can be used for obtaining absolute paleointensities of the Earth's magnetic field (see *Paleointensity from TRM*). Hence, all paleointensity methods require that the NRM be essentially pure TRM, or that the TRM can be readily isolated from the NRM. The paleointensity methods compare the NRM of each specimen to a new laboratory TRM produced in a known laboratory field, H_L . Because chemical and mineralogical alterations of specimens during laboratory heatings are common and often preclude reliable paleointensity determinations, the different paleointensity methods apply various pre- and post-heating tests to assess the extent of chemical changes on the remanence and paleointensity experiments. Several paleointensity methods use a single heating to above the specimen's highest Curie temperature to produce a total laboratory TRM.

However, because reaction rates increase with temperature, such a procedure tends to maximize chemical alterations.

The Thelliers' double heating method (Thellier and Thellier, 1959) was developed to diminish this problem by gradually heating the samples in steps from room temperature, T_R , to the highest blocking temperature. At each temperature step, $T_1 > T_R$, the samples are heated twice; to determine both the thermally demagnetized partial-NRM (PNRM) and the acquired partial-TRM (PTRM) between T_1 and T_R . The Thellier method depends on the additivity and independence of PTRMs acquired in different temperature intervals; that is, total-TRM = Σ PTRM. Also, it is assumed that the unknown paleointensity, H_U , was constant throughout the remanence acquisition process and that H_U and H_L are sufficiently small that both TRMs are linearly proportional to the imposed field. When these conditions are satisfied and in the absence of chemical changes on heating, H_U can be calculated from the ratio $\text{PNRM}(T_1, T_R)/\text{PTRM}(T_1, T_R) = H_U/H_L$ and should be the same for each temperature interval. Every temperature step gives an independent paleointensity value. These consistency checks, provided by several independent paleointensity estimates at the different temperatures, are the primary asset of the Thellier method. The data can be displayed on a PNRM versus PTRM plot (an Arai diagram), with data corresponding to the different temperature steps (Figure M107). Ideal behavior in the Thellier sense implies linear data with a slope equal to $-H_U/H_L$ (line A, Figure M107). The Thelliers' procedure is well suited to detect the onset of chemical alterations, which are more common at higher temperatures, and are often expressed as deviations from the ideal straight line. Another feature of the Thelliers' procedure is the PTRM check, where a PTRM is repeated at a lower temperature $T_P < T_1$. The PTRM check provides information about changes in the PTRM capacity of magnetic particles with $T_B \leq T_P$. For these reasons, the Thelliers' procedure is usually considered to be the most reliable paleointensity method.

High-temperature chemical alterations during the Thelliers' paleointensity procedure is one of the more common causes for failed or abbreviated paleointensity experiments. If the CRM is expressed as a

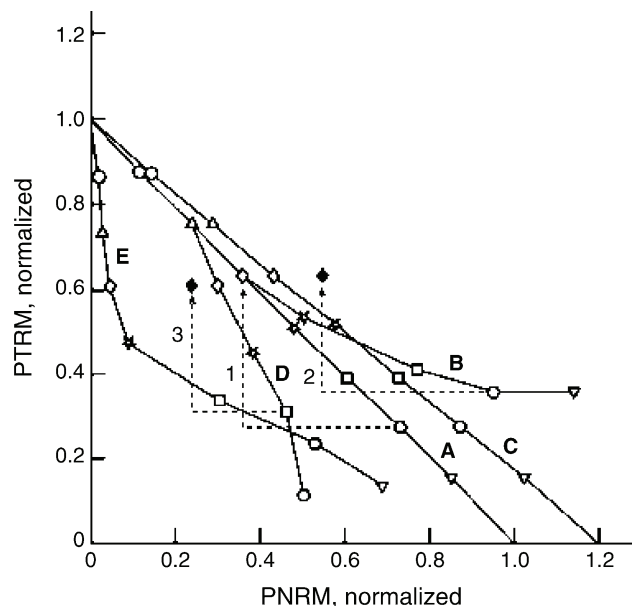


Figure M107 PNRM-PTRM diagrams for five hypothetical Thellier paleointensity experiments, A-E, discussed in the text. Like symbols indicate identical temperatures. Dashed lines 1, 2, and 3 indicate PTRM checks between the designated temperatures for experiments A, B, and D, respectively. Solid diamonds refer to unsuccessful PTRM checks.

greater PTRM capacity, which increases with temperature, the data will form a concave-up PNRM-PTRM plot (Figure M107, curve B). Provided the lower temperature data are linear and the PTRM checks show no evidence of alteration, then the lower temperature data can be used to calculate a paleointensity (Figure M107 curve B, points T1-T3). Recent studies suggest that at least 50% of the NRM should be used to obtain reliable results (e.g., Chauvin *et al.*, 2005).

CRM production that increases the PTRM capacity may result from precipitation of new magnetic particles or from unmixing of titanomagnetite grains to a more Fe-rich phase with higher saturation magnetization. Chemical modifications, which lead to higher PTRM, usually cause the PNRM-PTRM points to lie above the ideal line, and the calculated paleointensity will be lower than the actual paleofield. Alternatively, if the chemical alterations decrease the PTRM potential by destroying magnetic particles or by transforming them to a phase with lower intrinsic magnetic moments, then the PNRM-PTRM points will plot below the ideal line, with higher apparent paleointensities than the actual values (e.g., Figure M107, curve D and PTRM check 3). For a special case, where CRM acquisition grows linearly with temperature, the PNRM-PTRM plot might be linear (Figure M107, line C); however, these data plot above the ideal line (Figure M107, line A), and the calculated paleointensity would be lower than its actual value. This result emphasizes that linear PNRM-PTRM data are a *necessary but not sufficient* condition for obtaining reliable paleointensities.

The adverse effects of chemical alterations and CRM on paleointensity studies do not always arise from heatings in the laboratory. It is also possible that the NRM is not a pure TRM but contains a significant CRM component. Yamamoto *et al.* (2003) suggested that high temperature CRM contributes to the NRM of the Hawaiian 1960 lava, which results in higher than expected paleointensities. Alternatively, low-temperature hydrothermal alteration might produce CRM in new magnetic particles that contribute to the NRM. If these particles have not grown significantly beyond their blocking volumes, v_{AB} , they would be demagnetized at $T \geq T_A$, leading to rapid decrease of the NRM. This scenario might explain the precipitous diminution of the NRM observed for some basalts, with decreases on the order of 20% to more than 50% in the first few temperature steps of the Thellier experiment. When this decrease in NRM cannot be attributed to viscous remanence, it is possible that the NRM is augmented by CRM. It is no longer pure TRM.

CRM in Sedimentary Rocks

Oxidized Red Sediments

Red beds are a broad and loosely defined category of highly oxidized sediments with colors ranging from brown to purple, usually resulting from secondary fine particles of hematite, maghemite and/or ferric oxyhydroxide. The color is a complex function of the mineralogy, chemical composition and particle sizes of the iron oxides, as well as the impurity cations and their concentrations; however, for paleomagnetism, color is unimportant. Red sediments have been used extensively for paleomagnetism since the late 1940s, because they are widely distributed geographically and with respect to geologic time. In addition, the remanence of red sediments is often stable and sufficiently intense for paleomagnetic measurements, even with early-generation magnetometers. Already in the 1950s, it was deduced that low temperature oxidation was responsible for transforming the original magnetite to fine particles of hematite, maghemite, and/or goethite, which provide the pigment and CRM of red beds (Blackett, 1956).

Larson and Walker (1975) studied CRM development during early stages of red bed formation in late Cenozoic sediments; they showed that in their samples CRM occurred in several authigenic phases including hematite and goethite. The CRM, which obscured the original DRM, had formed over multiple polarity intervals, as indicated by different polarities in several generations of authigenic minerals. Complex multi-generation patterns of CRM, with several polarities within

single specimens, have also been observed in Paleozoic and Mesozoic red beds. The influence of the secondary CRM on the primary remanence depends on the relative stability and intensity of the CRM carriers as compared with the primary DRM. In many cases the DRM may have been entirely obliterated by diagenetic processes, and the CRM is the dominant characteristic remanence. However, there are examples of red beds, where the primary DRM in specularite hematite remains the characteristic remanence with respect to CRM (e.g., Collinson, 1974). Sometimes distinct CRM components can be isolated by thermal demagnetization, selective leaching in acids (e.g., Collinson, 1967) and removal of altered phases of sediment by selective destructive demagnetization (Larson, 1981).

During the past more than five decades, paleomagnetic studies of red sediments have contributed significantly to magnetostratigraphy, plate tectonics and rock magnetism. Many data of apparent polar-wander paths are from red beds, where it is assumed that the CRM was produced soon after deposition, so that the paleomagnetic pole accurately represents the depositional age of the sediments. Moreover, CRM is not subject to inclination shallowing, which often affects the primary DRM. The utility of red sediments for high resolution studies of the geomagnetic field and paleosecular variation is limited and depends on how pervasive the CRM is as compared to the primary DRM, the time lag between the CRM and initial DRM, and the duration of CRM production.

Non-red Sediments

Here we discuss CRM in a subset of non-red mostly carbonate sediments, whose remanence is usually much weaker than for red beds. The low remanence intensity of these sediments was a key reason that they were generally excluded from paleomagnetic investigations until the introduction in the early 1970s of cryogenic magnetometers, capable of measuring minute signals, down to the 10^{-9} – 10^{-10} Gauss range. Since then, there has been an explosion of studies of weakly magnetized non-red sediments. For example, it has been shown that the characteristic remanence of some early Paleozoic carbonate sequences was acquired in the late Paleozoic, hundreds of millions of years younger than their biostratigraphic ages (e.g., McCabe *et al.*, 1983). Magnetic extracts from these diagenetically altered sediments contained essentially pure magnetite particles, whose botryoidal and spheroidal forms have been used to infer their secondary origin, and they are thought to be responsible for the secondary characteristic remanence of these sediments. This conclusion has been buttressed by electron microscope observations (Figure M108) of in-situ authigenic magnetites in Paleozoic limestones (Suk *et al.*, 1993). In the absence of evidence of significant heating of these sediments, their remanence has been attributed to low temperature CRM in secondary magnetite, which might have been produced by “diagenetic alteration of preexisting iron sulfides (e.g., framboidal pyrites)” (McCabe *et al.*, 1983). For Miocene dolomites and limestones of the Monterey Formation, Hornafius (1984) concluded that the secondary remanence, presumably a low temperature CRM, resides in diagenetic magnetite produced by partial oxidation of pyrite upon the introduction of oxygenated meteoric groundwaters to the formation. CRM in some Paleozoic carbonates resides in hematite particles (e.g., Elmore *et al.*, 1985) produced by diagenetic dedolomitization, where oxidizing fluids with high calcium contents cause calcite replacing dolomite (McCabe and Elmore, 1989).

The presence of magnetite (and siderite, FeCO_3) in oil impregnated sediments was discovered by Bagin and Malumyan (1976), and Donovan *et al.* (1979) reported correspondence of near surface magnetic anomalies over an oil field with a higher concentration of magnetic minerals in the sediments. Paleomagnetic studies of remagnetized hydrocarbon-impregnated Paleozoic sediments (McCabe *et al.*, 1987; Benthien and Elmore, 1987) indicate a relationship between hydrocarbon migration and the precipitation of authigenic magnetite particles, carrying the secondary CRM. This scenario is supported by extracted magnetite spherules up to several tens of microns in diameter and

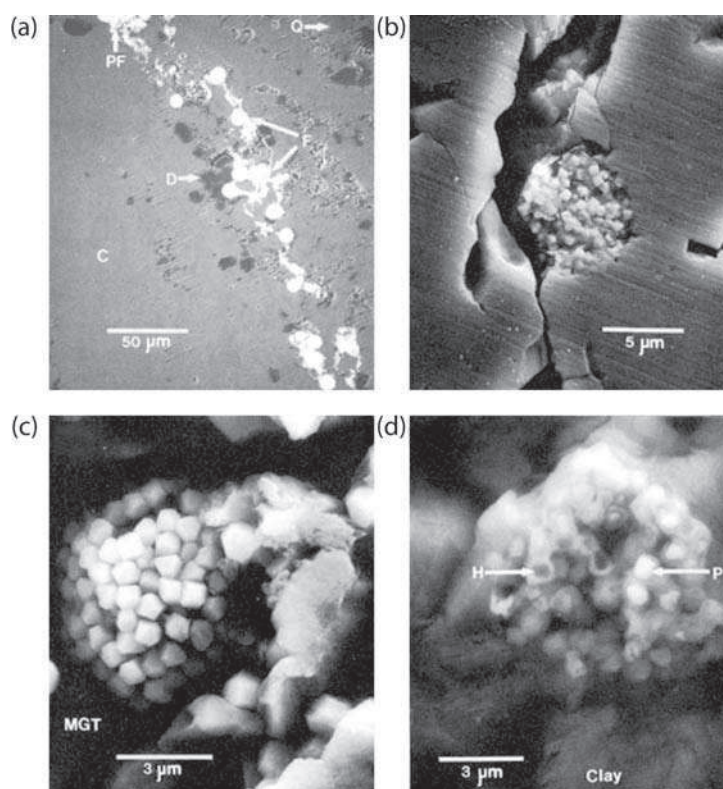


Figure M108 Scanning electron microscope images of pseudoframboidal magnetite in the New York carbonates. Symbols are MGT, magnetite; PF, pseudoframboid; F, framboid; P, pyrite; C, calcite; D, dolomite; Q, quartz; and H, hole. (a) Densely distributed framboids and pseudoframboids in a calcite matrix with occasional occurrence of dolomite and quartz; backscattered electron image (BEI). (b) Cross section of a pseudoframboid in a microcrack showing individual octahedral/cubo-octahedral crystals; secondary electron image (SEI). (c) A pseudoframboid in a microcrack showing almost perfect spherical shape (SEI). (d) An imperfectly spherical magnetite pseudoframboid in a void showing pyrite cores or voids within originally homogeneous pyrite crystals. Layered iron-rich clay minerals surround the grain (SEI) (from Suk *et al.*, 1993).

the presence of other authigenic textures in the sediments. All these studies suggest a net reduction of ferric ions in oxides, hydroxides and silicates, caused by the biodegradation (oxidation) of the hydrocarbons and the precipitation of more reduced iron oxide phases such as magnetite (Fe_3O_4), siderite (FeCO_3) and wustite (FeO). Of these, only magnetite is ferrimagnetic; hence it is responsible for the CRM and for being preferentially extracted during magnetic separations.

Rapidly deposited marine and lacustrine sediments are increasingly being used to study high-resolution behavior of the Earth's magnetic field, including secular variation and relative paleointensities. However, to accurately interpret the sedimentary record, geochemical processes that influence the magnetic signal must be understood. In anoxic and suboxic environments, bacterial sulfate reduction produces H_2S , which reacts with the detrital iron oxides to precipitate sulfide minerals (Bernier, 1970, 1984). An abundance of sulfate favors reactions that produce relatively more stable pyrite (FeS_2), which does not carry remanence. When the sulfate supply is more limited, the formation of ferrimagnetic pyrrhotite (Fe_7S_8) and/or greigite (Fe_3S_4) is preferred. Pyrrhotite formation is less common in sediments because it is thought to require $\text{pH} > 11$ (Garrels and Christ, 1965), which is outside the range of values measured in sedimentary pore waters. However, for extremely low sulfur activity, it is possible for pyrrhotite to form in sediments.

For anoxic sediments from the Gulf of California and suboxic hemipelagic muds from the Oregon continental slope with sedimentation rates exceeding 1m/kyr , Karlin and Levi (1983, 1985) documented very rapid, dramatic decreases with depth of the intensity of NRM

and artificial remanences, paralleled by downcore decrease in pore-water sulfate and systematic growth in solid sulfur, mainly as pyrite (Figure M109). In both environments, the remanence resides in fine-grain nearly pure magnetite. These data suggest that early oxidative decomposition of organic matter leads to chemical reduction of the ferrimagnetic minerals and other iron oxides, which are subsequently sulfidized and pyritized with depth. Changes in the remanence intensity and stability are consistent with selective dissolution of the smaller particles, causing downcore coarsening of the magnetic fraction. In these environments, there was no evidence for the formation of authigenic magnetitic minerals. In this example, there is no CRM formation; rather, the sediments experience chemical demagnetization via dissolution. The chemical processes cause substantial reduction of the remanence intensity, while the directions appear to be unaffected.

In other suboxic marine environments, characterized by lower sedimentation rates, on the order of centimeters/kyrs, CRM in authigenic magnetite particles accompanies oxidative decomposition of organic matter immediately above the Fe-reducing zone (Karlin *et al.*, 1987; Karlin, 1990). Some of the smaller authigenic magnetite particles are subsequently dissolved downcore on entering the zone of Fe-reduction (Figure M110). In the past approximately fifteen years, an increasing number of paleomagnetic studies have identified CRM in ferrimagnetic iron sulfides, pyrrhotite and greigite, in a variety of marine and lacustrine settings (e.g., Roberts and Turner, 1993; Reynolds *et al.*, 1999; Weaver *et al.*, 2002; Sagnotti *et al.*, 2005). While in many cases the Fe-sulfides are formed during early diagenesis upon initial burial, they can also result from later diagenesis, deeper in the sections.

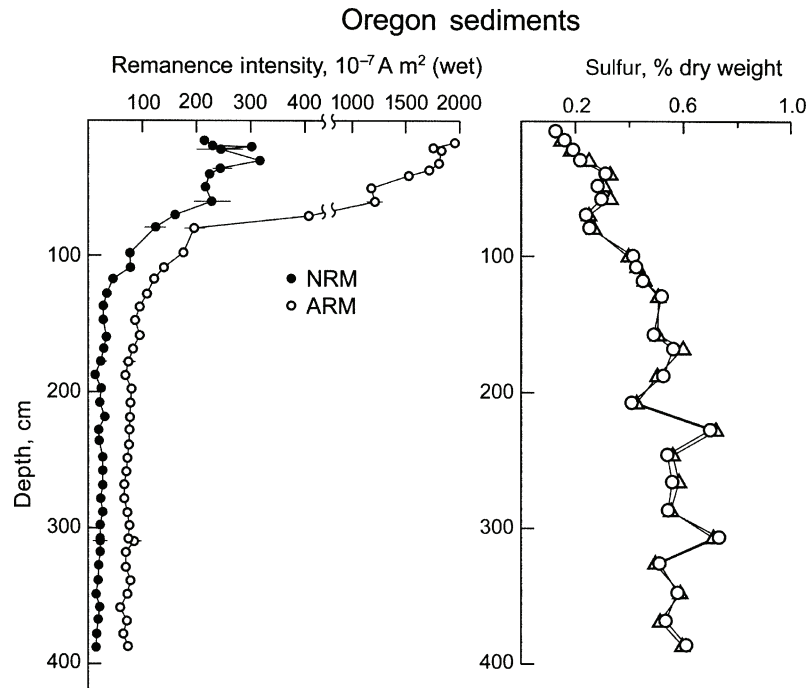


Figure M109 Downcore profiles of magnetic intensities and solid sulfur for Kasten core W7710-28, Oregon continental slope. The magnetization intensities were partially AF demagnetized at 15 mT. Solid sulfur concentrations of total (circles) and acid-insoluble (triangles) fractions on a sulfate-free basis, measured by X-ray fluorescence (modified from Karlin and Levi, 1983).

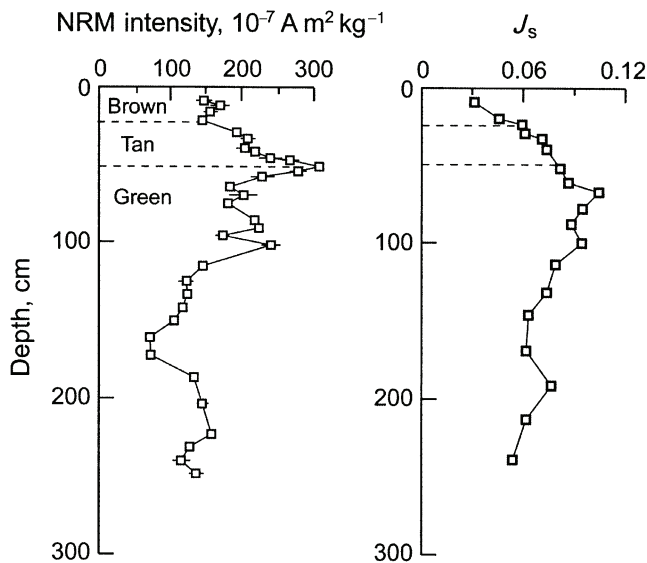


Figure M110 Downcore profiles of the NRM intensity, partially AF demagnetized at 20 mT, and the saturation magnetization for core TT11 from NE Pacific Ocean (modified from Karlin, 1990).

Recently, Roberts and Weaver (2005) described multiple mechanisms for CRM involving sedimentary greigite. The resolution of paleomagnetic time series would be compromised in sediments where CRM occurs in authigenic magnetic minerals, formed below the remanence lock-in depths. In light of this analysis it is possible that sediments discussed earlier in this section experienced multiple episodes of CRM. The first might arise from diagenesis of magnetite and other iron oxides to iron sulfides during initial burial. A later CRM

might be produced in magnetite due to diagenesis of iron sulfides or, possibly, in hematite due to dedolomitization in an oxidizing environment, depending on the sediment composition and prevailing geochemistry.

Concluding Remarks

This report on CRM is not exhaustive and reflects the interests, biases and limitations of the author. It is an update of a similar article written over fifteen years ago (Levi, 1989). In the future, as paleomagnetists address more difficult tectonic and geomagnetic questions, requiring data from structurally more complex, metamorphosed, and older formations, it will be increasingly likely that CRM will contribute to the NRM. Paleomagnetists have become more adept at isolating different remanence components, using detailed and varied demagnetization procedures. It is usually assumed that the most resistant remanence, whether with respect to increasing temperatures, alternating fields, or a particular leaching agent is also the primary component. However, CRM stabilities are highly variable, and this assumption is unlikely to be satisfied universally. During the past fifteen years, there has been progress in understanding several aspects of CRM, including (a) the recognition that even for some very young subaerial lavas the NRM may comprise a low-temperature CRM component, and (b) that in some active sedimentary environments, diagenesis leads to CRM in ferrimagnetic iron sulfides. A more comprehensive understanding of CRM is needed to assist paleomagnetists to interpret complex, often multicomponent, NRMs with probable CRM overprints. This goal would be advanced by conducting controlled field and laboratory CRM experiments to (1) recognize the varied geochemical environments that produce different magnetic minerals and their associated CRMs; (2) determine the ranges of magnetic and mineralogical stabilities with respect to different demagnetization procedures and for isolating different CRM components; and (3) develop procedures for identifying the timing and sequencing of multi-component CRMs.

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MAGNETIZATION, DEPOSITIONAL REMANENT

The magnetization acquisition processes in unconsolidated sediments have been long studied (e.g., Johnson *et al.*, 1948; King, 1955; Granar, 1958). The early studies showed that magnetic minerals in the sediments align along the ambient magnetic field during deposition through the water column. The magnetization resulting from the sedimentation process has been referred as depositional or detrital remanent magnetization (DRM). The magnetization acquisition process is still not well understood, and the role of the complex interplay of processes occurring during deposition, water-sediment interface processes, burial, and compaction, etc., require further analyses. The characteristics and stability of the remanent magnetization of unconsolidated sediments are determined by the composition, grain size, and shape of individual grains. During deposition in aqueous media, the magnetic particles are subject to the aligning force of the ambient magnetic field, plus the gravitational and dynamic forces. In tranquil conditions alignment of magnetic grains is relatively effective in depositional timescales, which are affected by Brownian forces. Nagata (1961) showed that equilibrium with the ambient magnetic field is attained in a scale of 1 s. Therefore, saturation magnetization should be expected in natural depositional systems, with the DRM intensity independent of the ambient magnetic field intensity. However, this is not the case, and the magnetization intensity is related to the intensity of the ambient magnetic field (Johnson *et al.*, 1948). The DRM intensities lower than saturation values have been related to misalignment effects of Brownian motion of submicron ferrimagnetic grains (Collinson, 1965; Stacey, 1972). Near the water-sediment interface, flow conditions may become relatively stable and simple by having laminar flow (Granar, 1958) but still the deposition process is complex. The interplay and characteristics of the bottom sediments result in a variety of fabrics in the deposited sediments.

In general, reorientation of magnetic minerals occurring after deposition and before consolidation of the sediment is referred as postdepositional DRM. Perhaps, the most notable distinction between depositional DRM and postdepositional DRM is the occurrence of the so-called inclination error present in depositional DRM (Johnson *et al.*, 1948; King, 1955; Granar, 1958). If I_H is the inclination of the ambient Earth's magnetic field at the time of sediment deposition, then the inclination of magnetization I_S can be expressed in terms of

$$I_S = \arctan(f \tan I_H)$$

where f is a factor that is determined experimentally. The inclination error has been ascribed to deposition of elongated grains with along-axis magnetizations tending to lie parallel to the sediment interface and deflecting the magnetization toward the horizontal plane. Laboratory experiments have been conducted to evaluate effects of the intensity of the ambient magnetic field, size and shapes of the magnetic grains, deposition on horizontal, inclined, and irregular surfaces, bottom currents, etc. (e.g., Johnson *et al.*, 1948; King, 1955; Rees, 1961). The inclination error has been observed in laboratory experiments, where

the angular difference can be as high as 20° (King, 1955), but it is smaller (5–10°) or absent in natural sediments.

Barton *et al.* (1980) studied the change with time in the DRM acquisition process of laboratory-deposited sediments and found that in less than 2 days, there was no appreciable inclination error. In natural conditions, postdepositional DRM presents no significant inclination error. One of the major differences between laboratory experiments and natural conditions is the deposition rate. The time taken for realignment has been estimated in a few years and is apparently related to the water content of the sediments. Verosub *et al.* (1979) experimentally reexamined the role of water content in acquisition of postdepositional DRM and suggested that small-scale shear-induced liquefaction is the main magnetization process. There are also several additional factors involved; for instance bottom water currents, changing water levels, presence of organic matter, biological activity (bioturbation), particle flocculation, floccule disaggregation, dewatering, etc.

In general, it appears that rapidly deposited sediments show inclination and bedding inclination errors, similar to those observed in laboratory experiments. Slowly deposited or high-porosity sediments show small or no inclination error. Postdepositional DRM will realign the magnetization direction; this process may occur in short timescales of days or months, but may occur in periods of years or decades following deposition (Tarling, 1983).

In addition to studies of secular variation and magnetostratigraphy in sedimentary sequences, there has also been much interest in determining relative paleointensities from sedimentary records (Tauxe, 1993). Long records of relative paleointensities have been derived from marine and lake sedimentary sequences, and results have been compared with volcanic records and other records. There have been also several attempts to examine the effects of depositional factors in the DRM intensity, including for instance the effects of clay mineralogy, electrical conductivity of sediments, pH and salinity (Lu *et al.*, 1990; Van Vreumingen, 1993; Katari and Tauxe, 2000). Katari and Bloxham (2001) examined the effects of sediment aggregate sizes on the DRM intensities, and proposed that intensity is related to viscous drag that produces misalignment of magnetic particle aggregates. They argue that interparticle attractions arising from electrostatic or van der Waals forces and/or biologically mediate flocculation results in formation of aggregates (which present a log-normal size distribution) preventing settling of individual smaller grains.

The depositional and postdepositional DRM in laboratory experiments and naturally deposited sediments have been intensively studied; nevertheless, further work is required to understand the complex interplay of processes and then develop magnetization acquisition models (e.g., Verosub, 1977; Tarling, 1983; Tauxe, 1993; Katari and Bloxham, 2001).

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