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Paleomagnetism of the Late Triassic Petrified Forest Formation, Chinle Group, western United States: Further evidence of "large" rotation of the Colorado Plateau

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Abstract. Paleomagnetic poles and magnetostratigraphy were obtained from the late Carnian and early Norian Petrified Forest Formation. The remanent magnetization characteristics of this variegated claystones formation are correlated closely with stratal color; red strata display univectorial magnetizations directed to the origin of orthogonal axes plots between 300° and 630°C, whereas progressively greater overlap in magnetization component stabilities is exhibited by purple, blue, and green strata. Rock-magnetic, petrographic, and microprobe investigations indicate that detrital titanohematite is the carrier of much of the characteristic magnetization, with a smaller contribution from magnetite in green and portions of the red strata. Eight magnetic polarity intervals were observed in three quarters of the formation (155 m); good correlation is observed with the coeval portion of the Newark Supergroup. Paleopoles from the late Carnian Blue Mesa and the early Norian Painted Desert Members of the Petrified Forest Formation are identical and agree well with other Chinle Group paleopoles from the Colorado Plateau; the tight cluster formed by all poles indicates that little or no apparent polar wander (APW) with respect to North America occurred during the Late Triassic (late Carnian through Rhaetian time). In addition, Chinle Group paleopoles derived from strata located on the North American (NA) craton also are tightly grouped and indicative of a paucity of significant APW through the Late Triassic. With that same observation from the coeval Newark Supergroup, three independent Late Triassic data sets from different tectonic settings in NA indicate minimal or no APW during more than 20 m.y. The combination of the vertebrate fauna and magnetostratigraphy within the Chinle Group provides a high degree of age control, allowing comparisons among coeval Colorado Plateau (CP) and NA craton paleopoles at about a 250-500 kyr level. CP Chinle Group paleopoles are displaced 9±3° clockwise from cratonic paleopoles. The Chinle Group is one of five sets of paleomagnetically sampled Mesozoic strata shown by magnetostratigraphy and/or biostratigraphy to have been deposited coevally on both the CP and the NA craton; paleopoles from the five sets indicate that CP paleopoles have been displaced a total of 10±3° relative to NA cratonic counterparts. Furthermore, the wealth of paleomagnetic and biostratigraphic data now available from both the Chinle and Newark strata demonstrates that nothing is sacred in NA tectonics: Newark Basin paleopoles are displaced 9±3° counterclockwise from western cratonic poles.

1. Introduction

The Chinle Group of the western United States (Figure 1) preserves the best Late Triassic terrestrial faunal and floral record of any global nonmarine sequence [*Lucas*, 1993]. Numerous fossils have been collected from Chinle strata. The Chinle Group consists of a large number of formations, whose definitions were prompted by numerous lateral changes in lithology. Figure 2 displays a representative sampling of formational names in the Chinle Group. This study focuses on the Petrified Forest Formation (Figure 2) because of its particularly abundant vertebrate fauna. Four faunachrons have been recognized in the Chinle Group [*Lucas and Hunt*, 1993]; the Petrified Forest Formation has a faunal assemblage typical of the Adamanian land-vertebrate faunachron. The formation is best exposed in the Petrified Forest National Park, where we magnetostratigraphically sampled about three quarters of it. Whenever possible, the paleomagnetic sampling maintained close proximity to the known fossil localities.

Previous paleomagnetic investigations of Chinle Group have shown that the strata preserve stable Late Triassic magnetizations and a record of numerous reversals of the geomagnetic field [*Reeve and Helsley*, 1972; *Reeve*, 1975; *Bazard and Butler*,

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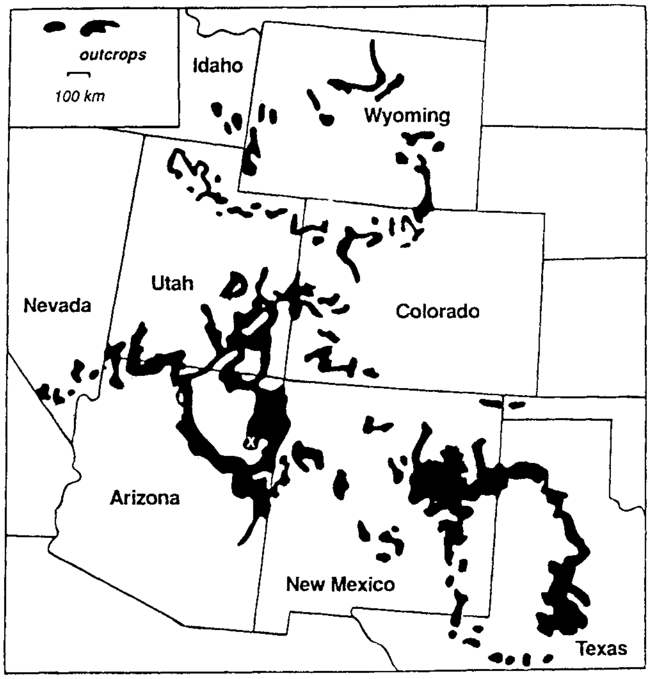

Figure 1. Distribution of outcrops of the Upper Triassic Chinle Group [after *Lucas*, 1993]. The white cross (X) near the Arizona-New Mexico border indicates the Petrified Forest Formation sampling localities.

1991; *Molina-Garza et al.*, 1993, 1996]. In fact, the presence of opposite polarity magnetization in the Chinle Group [*Reeve and Helsley*, 1972; *Reeve*, 1975] first indicated that something was amiss with the previous concept that wholly normal polarity record characterized this time, a result of the early paleomagnetic records obtained from the equivalent age Newark Supergroup red beds of the eastern United States [e.g., *Deboer*, 1968; *McIntosh et al.*, 1985].

In the past, correlation within the Chinle Group had been significantly hampered by the large number of facies and corresponding plethora of local names and by a lack of detailed biostratigraphy. Early Chinle magnetostratigraphic studies suffered from inaccurate lithostratigraphic correlations. *Reeve's* [1975] magnetostratigraphic sequences of Chinle strata appeared to yield conflicting results when correlated according to the accepted lithologic correlations at that time; at that time, the Redonda Formation of the Chinle Group in eastern New Mexico was thought to correlate with the Wingate Formation in Utah [*Stewart et al.*, 1972], and the "Chinle Formation" underlying the Redonda in eastern New Mexico was thought to be equivalent to the upper Chinle strata of Utah (then termed the Church Rock Formation; Figure 2a). The magnetic polarity stratigraphy definitely was at odds with these lithostratigraphic correlations, and no satisfactory explanation could be offered other than possible remagnetization [*Reeve*, 1975].

Despite the fact that *Reeve's* [1972; 1975] data were generated prior to modern laboratory and analytical procedures, the consistency of remanence polarity from sample to sample over discrete stratigraphic intervals [e.g., Figure 2 of *Reeve and Helsley*, 1972] and polarity stratigraphy that matched between nearby sections (Fig 2a) suggested that a stable, characteristic remanence had been isolated. The numerous polarity intervals, the approximately antipodal remanences, and polarity patterns which matched between distant outcrops all indicate that the Chinle remanence originated in the Late Triassic geomagnetic field near the time of deposition. These considerations led one of us (M. Steiner) to consider recorrelation of *Reeve's* [1972, 1975] sections by magnetostratigraphy, in disregard of the lithostratigraphy. *Reeve's* [1972, 1975] data were carefully reviewed and any samples interpreted as normal polarity by him, were examined in the light of their possibly being derived from demagnetization-resistant imprints of recent geomagnetic fields; any such data were excluded from a revised interpretation. Furthermore, unsampled intervals of >5 m and stratigraphic intervals in which the data did not yield a clear polarity signature were clearly identified as intervals lacking data in the revised magnetostratigraphic interpretation. Correlation based on the revised magnetostratigraphies (Figure 2b) was striking in that the stratal relationships indicated among Chinle Group formations were identical to those subsequently obtained through biostratigraphic and sequence stratigraphic correlations [*Lucas*, 1993]. Figure 2c illustrates how well these two correlation methods agree by displaying Reeve's [1972; 1975] data on the tetrapod faunachrons determined by Lucas [1993; *Lucas and Hunt* [1993]. A particular example of the success of the revised magnetostratigraphic correlation of Reeve's data is illustrated by the indicated coeval nature of the fluvial sandstones of the Church Rock Member of the Chinle Formation with the lacustrine shales of the Redonda Formation, which both also have the same biostratigraphy [*Steiner and Lucas*, 1995].

The early Chinle studies by *Reeve* concentrated on the upper formations of the Chinle Group. *Molina-Garza et al.* [1993; 1996] have investigated the magnetostratigraphy of formations of many of the formations in the lower part of the Group, specifically the Santa Rosa, Garita Creek, and Trujillo Formations (Figure 2). Yet the formation of the Chinle Group bearing the most important fossil record, the medial Petrified Forest Formation of western New Mexico-eastern Arizona and southern Utah (Figure 2c) had never been studied magnetostratigraphically. This stratigraphic interval is almost entirely claystone and thus is difficult to sample by conventional paleomagnetic techniques; nevertheless, the abundance and importance of its fauna motivated us to undertake a detailed magnetostratigraphic collection of it.

2. Geology and Sample Collection

The Petrified Forest Formation of the Chinle Group consists of variegated, bentonitic claystones with minor interbedded sandstones. It has been divided into three members [*Lucas*, 1993], two thick claystone intervals (the Blue Mesa and Painted Desert Members) separated by a laterally persistent sandstone (Sonsela Member; Figure 2c). The multihued claystones are a hallmark of the formation and are responsible for the name Painted Desert, used to describe this region of the western United States. The Blue Mesa Member of the Petrified Forest Formation consists dominantly of green, white, blue, purple, and a few red claystones; the Painted Desert Member is largely red claystones, but with variegated colors in the lower strata. The dividing Sonsela Member is a white, medium- to coarse- grained sandstone and conglomeratic sandstone. Vertebrate biochronology [*Lucas*, 1993] and palynostratigraphy [*Litwin et al.*, 1991] indicate that the Sonsela Member also marks the position of the boundary between the Late Triassic Carnian and Norian Stages.

We sampled the Blue Mesa Member at its type locality [*Lucas*, 1993], Blue Mesa (250.1°E, 35.0°N), in the Petrified

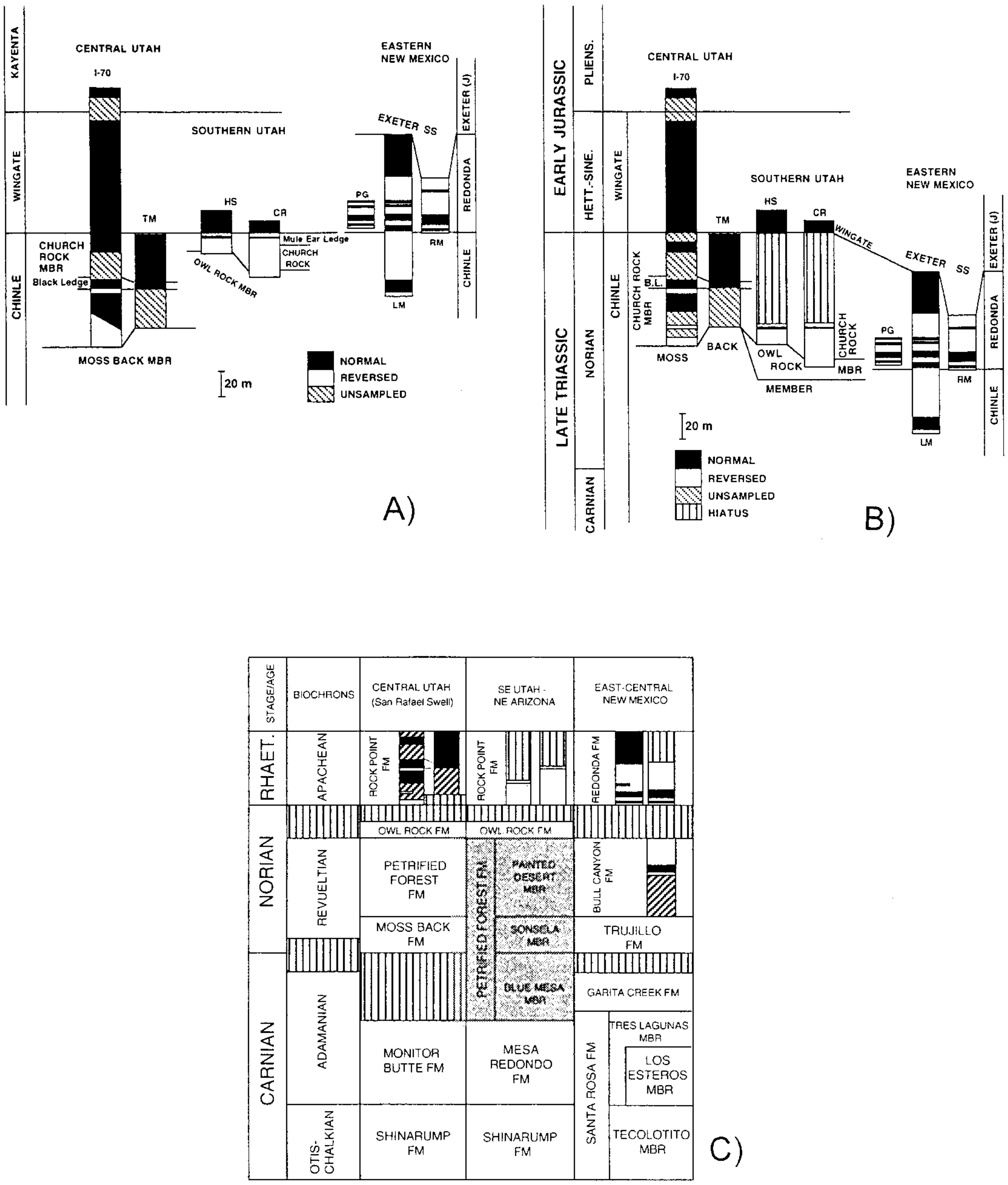

Figure 2. Correlation of *Reeve*'s [1975] Chinle Group data based on magnetostratigraphy. Vertical lines represent hiatuses; diagonal pattern represents unsampled intervals or those with non-diagnostic data. (a) Reeve's original magnetostratigraphic correlation based on the accepted lithostratigraphy of the time; (b) Reeve's data re-correlated here on the basis of magnetic polarity; (c) an overlay of the magnetostratigraphy of Reeve's data onto the biostratigraphic correlation of the Chinle Group [Lucas, 1993]. (Note that in Utah, the term "Petrified Forest Formation" is restricted to the early Norian time interval, whereas the Petrified Forest Formation studied in this investigation of eastern Arizona strata extends from the late Carnian through the early Norian (stratigraphy shown in Figure 9a)).

Forest National Park. Sampling began at the lowest stratigraphic exposures which occur west of Blue Mesa and west of the paved park road. Sampling was carried up-section to the exposures adjacent to the road and continued up-section in exposures on the east side of the road until reaching the base of Blue Mesa itself; the remainder of the member was sampled up the mesa slopes, terminating at the basal sandstone of the Sonsela Member. The Sonsela Member itself was not sampled at Blue Mesa because of its coarse grain size there.

A second magnetostratigraphic section ~21 km north northwest, in the Lacey Point area of Lithodendron Wash, continued the magnetostratigraphic sampling up-section; these exposures constitute the type section of the lower part of the Painted Desert Member [*Lucas*, 1993]. Sampling in the Lacy Point area began below the Sonsela Member in the uppermost Blue Mesa Member, included some claystone intercalations within the lower Sonsela sandstone, skipped the coarse sandstone of the upper Sonsela, and continued through the lower half of the Painted Desert Member to terminate ~2 m below the tuffaceous Black Forest Bed.

All colors of strata in this multihued formation were sampled. Blue Mesa Member samples consisted of white, green, blue, purple and red siltstones and claystones and a few pink, white, or green sandstones. Painted Desert Member samples contain similar colors, but a large proportion were red siltstones and claystones. A number of the interspersed thin sandstones also were sampled; these had colors of white, brown, light red, or pale purple. Sampling sites were spaced 0.3 m to several meters apart except for intervals in which the lithology would not support intact samples or where lack of exposure prevented sampling. Three to five samples were obtained at each site. The lower member yielded 140 samples, collected over 77 m, and 153 samples were collected over 78 m in the upper member. All strata are flat-lying.

3. Paleomagnetic Data

3.1. Measurement/Demagnetization

The initially measured natural remanent magnetization (NRM) of red claystones mostly exhibited directions concentrated around either the normal or reversed Late Triassic field directions for the Colorado Plateau, although some were arrayed along great circles whose endpoints were the Triassic directions and the present field direction for this site. Red claystone NRM intensities varied between 2×10^{-2} and 2×10^{-4} A/m. Nonred (white, green, blue, purple) claystone NRM directions were scattered over the lower hemisphere, but none matched recent (present or axial dipole) field directions. NRM intensities of nonred claystones were between 3×10^{-3} and 3×10^{-4} A/m.

Both alternating field (AF) and thermal demagnetization techniques were applied to representatives of all of lithologies collected. Paired AF and thermal demagnetizations were performed on 50 samples from the same stratigraphic horizons. White, green, blue, purple, mottled red-purple, red-green, and red-white claystones were AF demagnetized in fields of 2.5 to 60 mT. Pilot thermal demagnetization employed temperatures of 160/200/235/300/350/400/440/480/510/530/550/560/570/600/620/640/650/660°C.

With the exception of green claystone intervals, AF demagnetization did not induce significant separation of remanence components, neither appreciably reducing remanent intensities nor effecting any change in directions. Red clay- and siltstones and purple and blue claystones exhibited ~25% reduction in remanent intensities, but no change in remanence directions. Thermal demagnetization thus provided the best separation of components of magnetization in these strata; optimum treatment temperatures differed according to sample color, but included about 12 of the steps used in the pilot study. The pilot samples showed red strata to be largely unaffected by temperatures below 300°C, hence general treatment of red strata began at 300°C. Heatings for all other samples began at 200°C.

Commonly, a present-day component was removed from most samples, removed below ~300°C. Thereafter, a large percentage of samples of all colors exhibited linear decay of remanence in part or all of the temperature spectrum between 300° or 400°C and 630°C. At higher temperatures, some samples continued to display directions similar to the ≤630°C, whereas others displayed widely divergent directions; linear decay behavior generally ended above 630°C. The remanence decaying linearly in the 300° to 630°C range was considered characteristic of a sample and hereafter is referred to as the "characteristic remanence". Characteristic magnetizations were defined by least squares fitting of lines to the linear portion of demagnetization trends [*Kirschvink*, 1980].

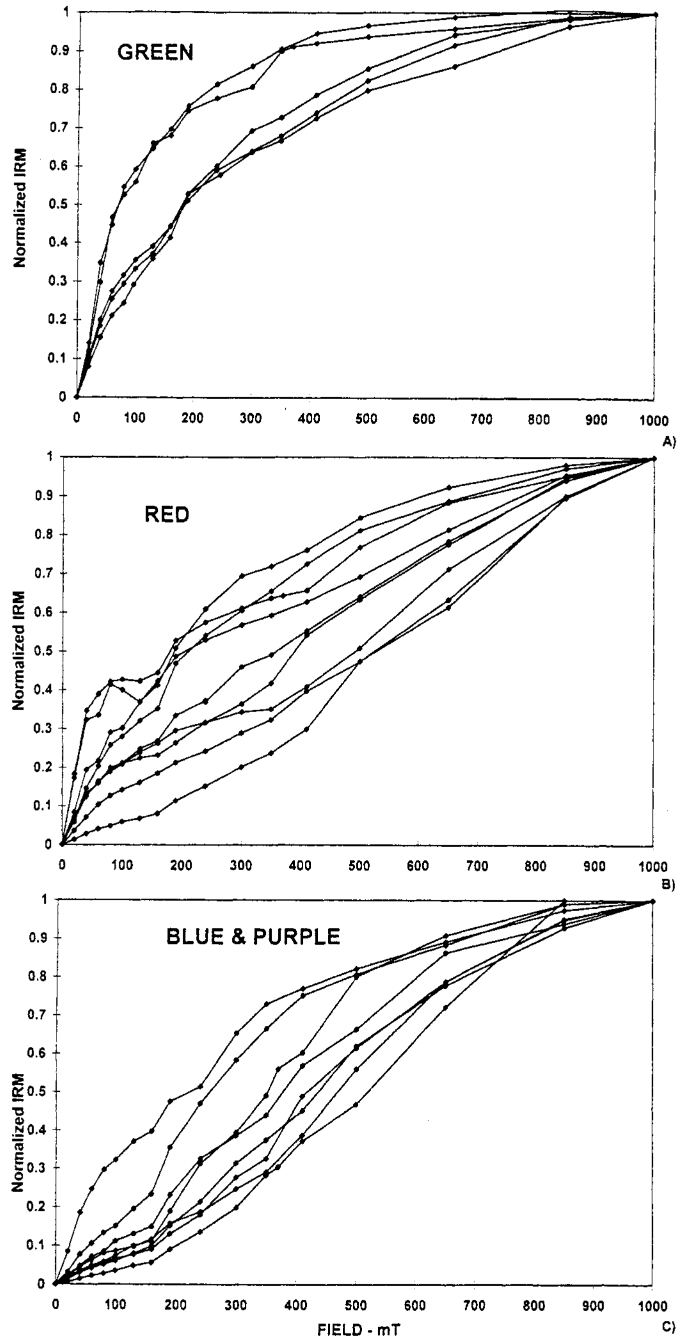

Figure 3. Isothermal remanent magnetization (IRM) acquisition in representative samples of green, red, purple, and blue claystones. (a) The two highest intensity curves from green claystones; the three of slightly lesser intensity from green, red and green, and purple and green samples, respectively. (b) Red claystones. (c) Purple and blue claystones (plotted on the same plot because they show the same ranges of acquisition characteristics).

3.2. Investigation of Magnetic Carriers

Acquisition of isothermal remanent magnetization (IRM) was investigated in representative examples of red, purple, blue, green, and mixed color claystones and one sandstone; 28 samples were subjected to IRM acquisition in fields from 10 to 1000 mT. The IRM acquisition curves (Figure 3) indicate a mixture of remanence carriers; several curves from the green sediments clearly indicate the dominant presence of magnetite or another spinel-structured mineral such as maghemite (Figure 3a), whereas other green and some red samples indicate the presence of both a spinel-structured mineral and a hematite in the same samples. Finally, a number of the blue, purple and red samples indicate that a hematite is the dominant magnetic mineral (Figures 3b and 3c).

Block samples of red, purple, blue and green claystones were pulverized and their iron-oxide minerals separated magnetically for the purpose of performing thermomagnetic measurements and petrographic and microprobe analyses. However, only the red claystones yielded sufficient magnetic material for thermomagnetic measurement, and even that amount was quite small. Consequently, only a single thermomagnetic curve was acquired. The magnetic extract from the red claystones displays nearly complete loss of magnetization by 580°C, but a small tail of magnetization persists out to 630°C, after which no other magnetization is visible. The cooling curve exhibits a twofold decrease in room temperature magnetization; these characteristics suggest that a Ti-magnetite series mineral dominated the extract signature, accompanied by either a Ti-hematite or a maghemite.

Seven polished thin sections and one of the four grain mounts of separated magnetic materials were examined petrographically in reflected light. Approximately three-fourths of the opaque oxide grains observed are hematites that display evidence of having been derived from magnetite. These grains are martite, hematite displaying the former 111 crystallographic planes of pseudomophed magnetite. Martite grains are of three main types; one is composed entirely hematite, but with crystal growth along the former 111 planes making these visible. The second form is the most common, one in which a 111-plane latticework is formed of hematite enclosing empty spaces or spaces filled with granular or amorphous TiO_2. A third form also is common, consisting of grains which are dominantly hematite, but contain small exolved ilmenite lamellae which define former 111 planes of magnetite.

In addition, most of the polished sections each exhibit a small number of grains of a fourth type; these are grains composed of a Ti-magnetite host with laths of hematite or ilmenite along the 111 crystallographic planes. In these grains the optically visible magnetite occurs as 2-5 micron triangular or rectangular regions. The presence of spinel in the magnetite demonstrates that of some of these grains formed by oxy-exolution of magnetite at high temperatures (>600°C [*Buddington and Lindsley*, 1964]), that is, prior to deposition. A minor amount of high-temperature-exolved grains of hemoilmenite (hematite with exolved ilmenite lamellae) also is observed, approximately one to five such grains per thin section.

Yet one other type of grain was observed. Numerous grains with a reflectance brighter than magnetite display curvilinear cracks; the crack character and patterns are identical to those formed in the early stages of maghemitization of magnetite. Maghemite forms in low-temperature aqueous environments by oxidation of magnetite, the consequence of migration of Fe cations out of the crystal lattice and resulting in lattice vacancies. The lattice shrinks from the loss of cations, and the shrinkage is expressed physically as curvilinear cracks in the grains. The cracks observed in the Petrified Forest Formation polished sections are identical to ones observed in oceanic basalts [*Ade-Hall et al.*, 1976; *Johnson and Hall*, 1978; *Steiner*, 1981]. The grains with curvilinear cracks are present in many of the thin sections; they have a relatively high reflectance and anisotropy under crossed nichols, suggesting a hemo-ilmenite composition now. It is thus speculated that these grains are hematite pseudomorphs of maghemite. The pseudomorphing of magnetite (martitization) is common in red bed strata [e.g., *Steiner*, 1983]; perhaps under appropriate conditions, maghemite can be similarly pseudomorphed. If these grains are pseudomorphs of maghemite, they may suggest an early wet depositional environment in the Petrified Forest Formation which subsequently turned drier subsequent to maghemite formation, allowing the maghemite to be further oxidized to hematite.

Most of the opaque oxide grains observed, including the magnetite grains, are subrounded to well rounded. Two of the sandstone thin sections display laminae of heavy minerals; in these cases, well-rounded iron oxide grains of oxy-exolved magnetite, hematite, and hematite-ilmenite are aligned along several bedding planes in each section.

Microprobe analyses were carried out on 35 iron-oxide grains in three grain mounts (from red, purple, and green claystones) and three thin sections. Grains normally were analyzed at three spots on each grain. Twenty eight grains, 80%, reside in the hematite-ilmenite solid solution series; two of the 35 grains had a titanomagnetite composition, and five others had ferrian-rutile compositions. The hemo-ilmenite grains possess a range of titanium contents, from 10% to 36% Ti substituted for Fe in the Fe_2O_3 lattice.

3.3. Demagnetization

3.3.1. Red siltstones and claystones. The remanence of all red samples displays linear decay to the origin of orthogonal axes plots as a consequence of thermal treatment. Generally all red samples exhibit this linear remanence decay over the entire 300° to 630°C temperature range (Figure 4a, 4b). Least-squares lines fit to the linearly decaying remanence all possess quite small errors (maximum angular deviation or MAD values), commonly 1-3° and always ≤5°.

Because magnetic separation generally did not yield adequate material for thermomagnetic analyses, remanence losses relative to temperature were examined. Two types of demagnetization response were discerned. The Blue Mesa Member red samples showed rather uniform decay of remanent intensities between 230° and 600°C, but with increased loss between 350°C and 600-630°C, above which large, abrupt intensity losses occurred (Figure 4c); some samples displayed increases in intensity between 600° and 630°C. Red samples of the overlying Painted Desert Member exhibit two types of remanence decay with increasing temperature; one half exhibits rather continuous loss of remanence to temperatures of 630°C, with no increased losses at the high values (Figure 4d). The other half of the Painted Desert red samples behave like the Blue Mesa samples, except that the abrupt remanence loss above 600-630°C is not observed (Figure 4e).

3.3.2. Green strata. Generally, NRM intensities were weak; however, green-colored strata were the exception to the general failure of AF demagnetization to resolve remanence components in the Petrified Forest Formation. AF demagnetization clearly resolved a Late Triassic component in samples from each of the green claystone intervals (Figures 5a and 5c-5d); sometimes thermal demagnetization in the same interval revealed the same component but commonly did not (Figure 5b). Although there were a few exceptions, thermal demagnetization generally failed to separate remanence components in green intervals. In addition to the solely green-colored claystones, a number of mixed, either mottled or interlayered, green-purple and green-red samples

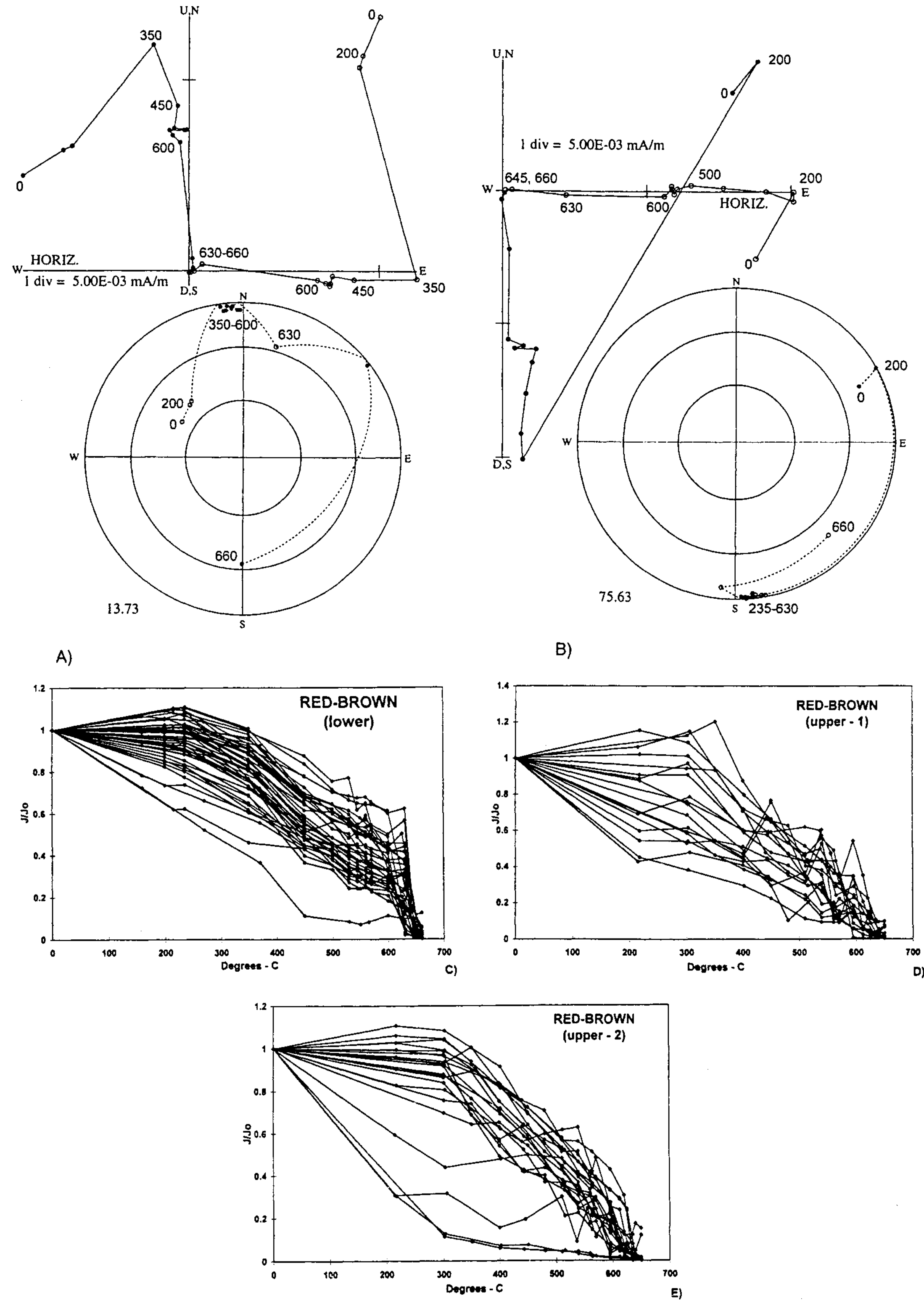

Figure 4. Thermal demagnetization of typical red claystones in the Petrified Forest Formation. Equal area stereographic plots show lower (upper) hemisphere directions as solid (open) circles. Orthogonal axes plots show declination as solid symbols, inclination as open symbols: (a) reversed and (b) normal polarity; (c) Normalized remanence decay with temperature for the red claystones of the lower, Blue Mesa Member; (d) normalized remanence decay for part of the red claystones of the upper, Painted Desert Member; (e) remanence decay for the rest of the red claystones of the upper member.

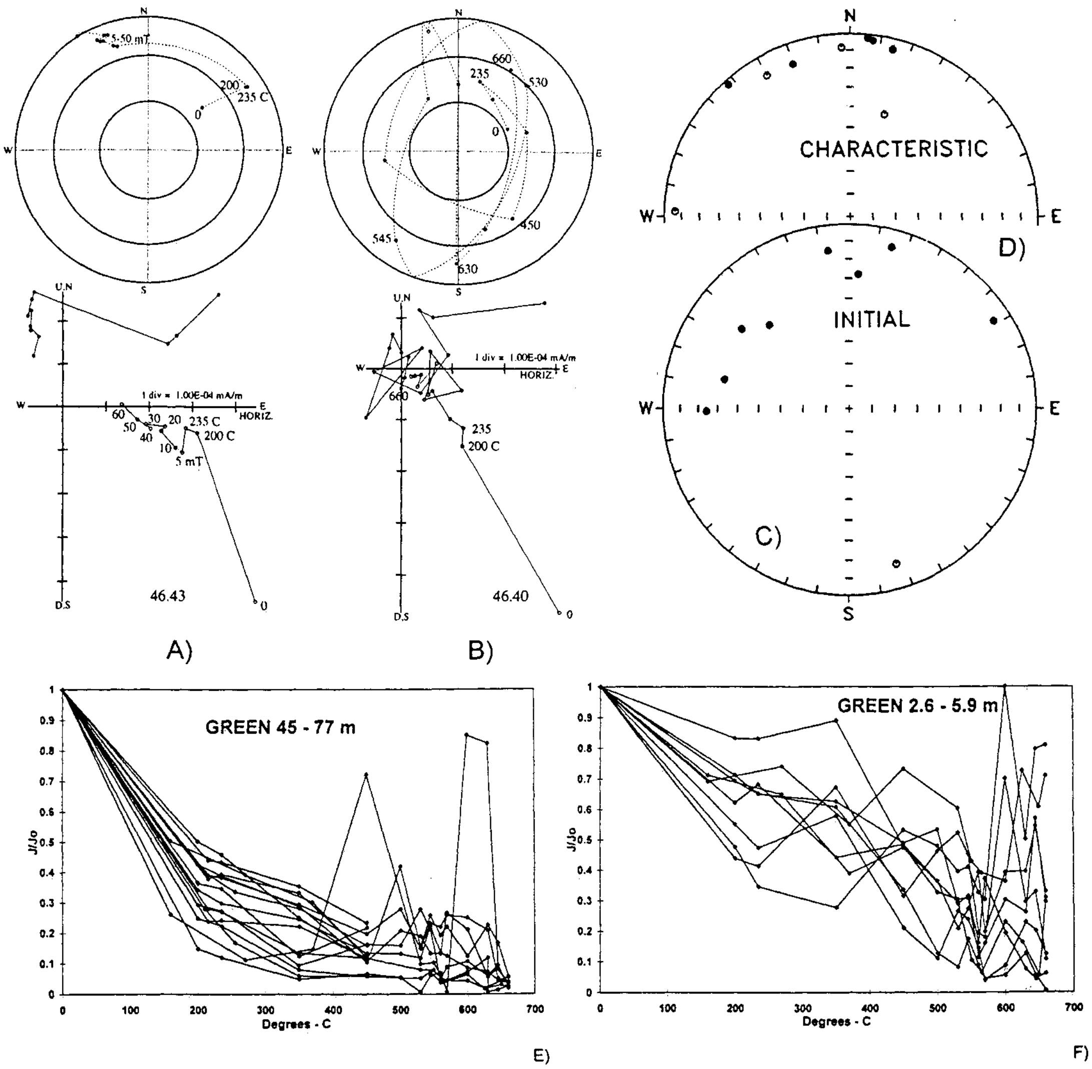

Figure 5. Demagnetization responses of green claystones: (a) AF demagnetization and (b) thermal demagnetization of samples from the same stratigraphic horizon (green unit 14 of *Lucas* [1993]); (c) and (d) initial (NRM) and characteristic directions for green claystones demagnetized by AF. E) Normalized remanence decay with temperature for the green claystones of the Petrified Forest Formation (Blue Mesa Member, units 17, 14, and 3 of *Lucas* [1993]; F) Normalized remanence decay for the Lucas's units 19-20 in the Blue Mesa Member at the Blue Mesa locality. All symbols as in Figure 4.

were investigated. All such color associations displayed poorer remanence preservation, lack of linearity during demagnetization, or diminished number of demagnetization steps over which linearity is displayed, than samples of that color which had no admixture of green color.

IRM acquisition in green claystones clearly indicated the presence of magnetite or another spinel-structured (maghemite) iron oxide. The two curves displaying the strongest remanence acquisition in Figure 3a are green claystones from two different stratigraphic intervals. The three lower-acquisition curves are samples of (1) a third green claystone interval, (2) a mixed red and green claystone interval, and (3) a mixed purple and green claystone. All curves indicate the presence of a cubic-structured magnetic mineral (magnetite or maghemite); the three of lower intensity indicate a mixture of cubic and rhombehedral (hematite) iron oxides, consistent, in two cases, with their mixed green and reddish color characteristics. Green claystone remanence loss with increasing thermal demagnetization shows one characteristic that is unique to the green-colored samples; most experience dramatic remanence losses by 200^{oC} (Figures 5a, 5b, and 5c).

One interval of green claystone differs from the rest; samples from the interval directly underlying the Sonsela Sandstone Member at Blue Mesa retain more remanent magnetization after heating to 200°C. A second characteristic of these samples is that a number of them exhibit pronounced remanence loss at 550-570°C (Figure 5d).

3.3.3. Blue and purple strata. Thermal demagnetization was reasonably effective at separating the remanence components in claystones of these colors but less so than in red claystones. Purple claystones from both the lower and upper members of the formation display the same demagnetization behaviors, as do the

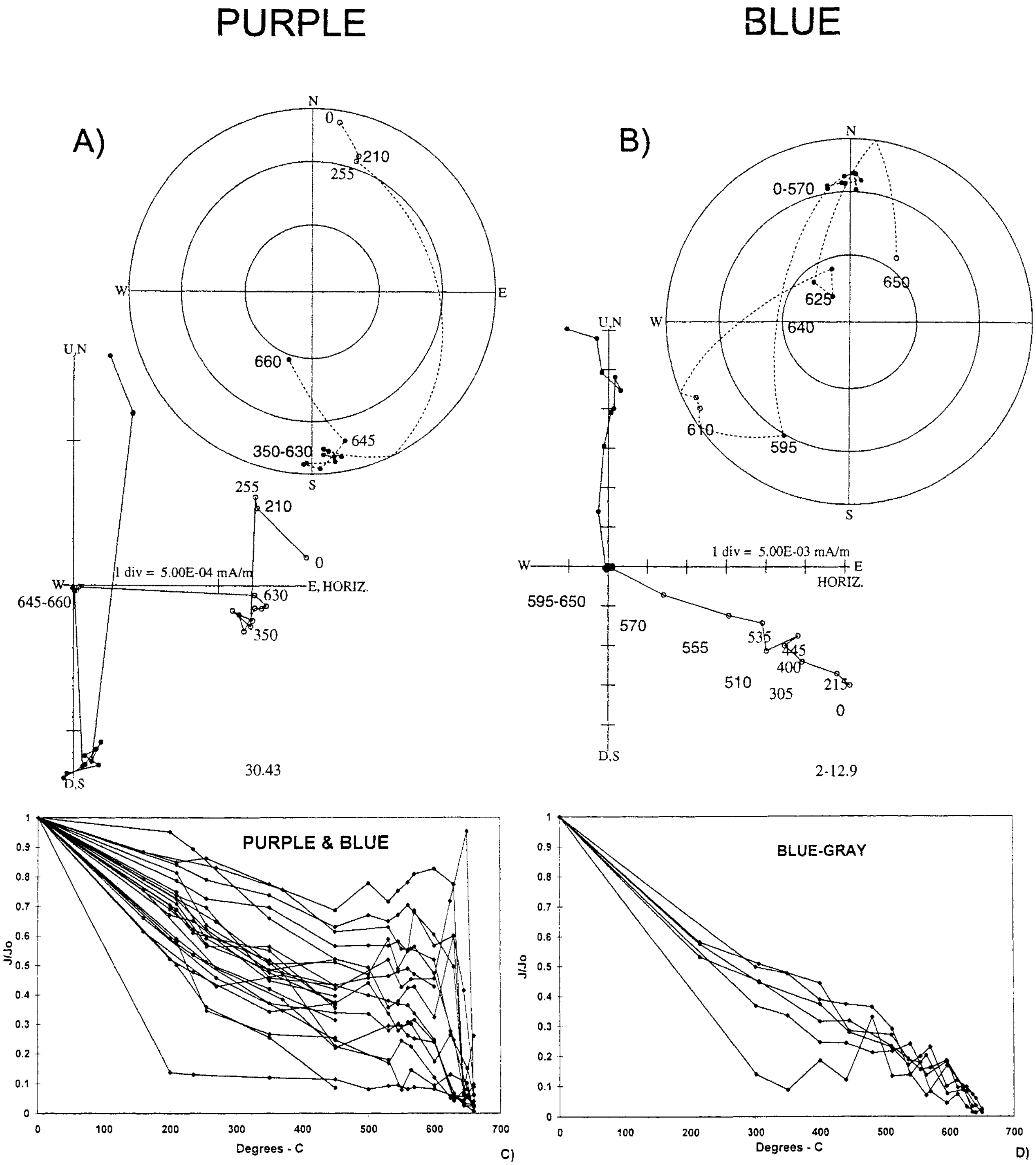

Figure 6. Thermal demagnetization behavior of (a) typical purple and (b) typical blue claystones from the Petrified Forest Formation. Symbols are as in Figure 4. (c) Normalized remanence decay with temperature for blue and purple claystones; and (d) normalized remanence decay for Lucas's units 19-20 of the Blue Mesa Member at the Lithodrendron Wash locality.

blue claystones. A typical purple claystone demagnetization response is shown in Figure 6a, and a typical blue claystone response in Figure 6b.

Purple and blue claystones are being discussed together because a color continuum exists between them, more so than between purple and red claystones. Remanent intensity decay with temperature is about the same in both purple and blue claystones (Figure 6c). For most, remanence loss is uniform until 450°C, after which sample intensity appear to plateau or to increase slightly. Some changes in remanence, commonly increases, are observed above 550°C; most claystones retain some remanence up to 660°C. Some of the blue claystones contain a hue of gray and behave significantly differently from entirely blue claystones (Figure 6d); they exhibit large remanence losses by 200°C, reminiscent of the green claystones, after which remanence loss is essentially continuous from 200° to 660°C, with no 450° and 600°C intensity plateaus or changes.

IRM acquisition in purple and blue claystones (Figure 3c) indicates mainly hematite as the magnetic mineral present; one curve out of eight suggests the presence of some cubic-structured magnetic mineral.

3.3.4. Sandstones. Nine of the Petrified Forest Formation sandstones intervals were sampled. Nearly all display multiple remanence components with significantly overlapping stability ranges. Neither AF nor thermal demagnetization techniques was effective in separating the components, and fairly ragged orthogonal axes diagrams characterize most sandstone samples. Sandstone color appeared to be less correlated to the degree of overlap of remanence component stabilities than did the sand grain size, although the sandstone which showed the most coherent within-body directions (PDF3 [*Lucas*, 1993]) has a pale red to purple color. A few of the samples from PDF3 exhibited fairly linear decay of remanence between 300° and 630°C, but many others exhibited nonlinear decay. Sandstone interval PDF1 [*Lucas*, 1993] contained laminae of heavy minerals, and the three samples possessing heavy mineral laminae displayed somewhat linear remanence decay, again in the 300-630°C temperature range. The overall failure of the sandstones to retain a stable remanence may speak to their hydrogeologic setting as thin sandstone intervals encased in thick claystone sequences.

4. Discussion of Results

4.1. Magnetization-Color Relationships and Timing of Magnetization

A distinct relationship is evident between the degree of overlap among unblocking temperatures of remanence components and the color of the strata. Least overlap in unblocking temperatures is observed in red strata, greater in purple strata, considerable in blue, and commonly nearly complete in green claystones. This relationship was examined in some detail to assess whether it had significance for timing of remanence acquisition, with the obvious implications for the magnetostratigraphy. The colors simply may be indicative of different depositional environments, or certain colors may have been developed later, in which case the characteristic remanence of those stratigraphic intervals also may represent later remanence acquisition.

The stratal colors most likely reflect different oxidation states of iron and thus provide an indication of the last pH environment which the sediment experienced. Green coloration commonly is a product of reducing environments, and Fe^{2+} is soluble in reducing environments. Hence, detrital iron oxides will readily go into solution in depositional settings which generate green sediments; Fe^{2+} will dissolve immediately, whereas Fe^{3+} will be converted to Fe^{2+} and then also will go into solution. Green strata therefore would be expected to have poorer remanence preservation, which is what is observed generally in paleomagnetism and specifically in this formation. Detrital iron oxides are in relative equilibrium in oxidizing environments and therefore far more likely to be preserved; consequently, a better defined remanent magnetic signature is to be expected in more oxidized sedimentary rocks, and all red strata of this study have very well-defined magnetizations. Blue and purple rocks probably result from a combination of oxidization and reduction in their environmental histories, and the mixed history may account for remanence characteristics intermediate between well and poorly defined.

The Late Triassic of western North America was dominated by fluvial, with some amount of lacustrine, sedimentation. The climate during deposition was strongly seasonal due to the megamonsoonal climate of Pangea [*Stewart et al.*, 1972; *Dubiel*, 1989; *Lucas*, 1993; *Parrish*, 1993]. Seasonal alternately wet and dry environments resulted in highly oxidizing conditions during the arid part of the years. Some permanently moist environments probably existed in the form of lakes, swamps, and nonephemeral stream channels and may have served as the depositional sites for green strata. The more oxygenated stream bank and interfluve environments probably hosted the red, purple, and blue sedimentary deposits, although drying of previously water-saturated regions such as ephemeral lakes and pools and stream channels probably also contributed to the generation of the blue, purple, and red deposits. The paucity of IRM evidence for a cubic magnetic mineral distinguishes purple and blue claystones from red ones; combined with the greater overlap of unblocking temperatures compared to red samples, these two characteristics suggest greater dissolution of spinel carriers in purple-blue relative to red sediments.

Two formation-wide remanence characteristics are notable. No one polarity of remanence is associated with any specific color of strata, suggesting that the characteristic remanence of the various colors of strata is not a result of discrete chemical oxidation or reduction events at some time long subsequent to deposition. Most significant is the fact that linear demagnetization behavior, thus the stability of the characteristic remanence, in all colors of strata is restricted to temperatures below 630°C. Most remanent intensity is lost from red, blue, and purple strata by 600° or 630°C (Figures 4c-e, and 6c-d). Directions commonly change markedly above 630°C. Even the reddest samples do not exhibit any significant amount of remanence at unblocking temperatures above 630°C. This unblocking temperature ceiling in the Petrified Forest Formation is surprising because red sedimentary rocks commonly lose much of their remanent magnetization between 660 and 680°C, the unblocking temperature of fine-grained detrital hematite. The fact that remanence loss in this temperature range is observed in different colors of strata suggests a mineralogical origin, stemming from proximity to a Curie, Neel, or inversion temperature.

An unblocking temperature of 600-630°C is not one normally associated with hematite and could represent remanence unblocking at either the Curie (T_c) or an inversion (T_{inv}) temperature of maghemite. The maghemite T_c is approximately 645°C [*Ozdemir and Banerjee*, 1984], although small amounts of impurities in the maghemite lattice act to stabilize the cation-deficient structure. (Michel et al. [1951] observed maghemite with 7% aluminum in the lattice to have a T_c of 591°C.)

A number of aspects of the Petrified Forest magnetization indicate some amount of magnetite is present today. (1) Isothermal remanence acquisition indicates a spinel phase in green and red strata. (2) AF demagnetization successfully revealed a Late Triassic remanence direction in green strata, indicating that some of the formation remanence is carried by magnetite or maghemite. (3) Petrographic observations demonstrate that some amount of magnetite is present in these sedimentary rocks. 4) Significant remanence loss between 530° and 580°C is displayed by a small number of samples; although such behavior has been observed in all colors of strata, it occurs mainly in the green (Figure 5f) and red strata (Figure 4d). These temperatures suggest remanence unblocking in nearly pure to slightly Ti-rich magnetite. Although rock magnetic and petrographic observations indicate only a comparatively minor amount of magnetite in these strata, thus only a minor role in the characteristic remanence, the importance of the magnetite which is present is that it has to have been a detrital component. Neither the oxidizing environments responsible for the red sediments nor the reducing environments inferred for the green sediments, would have permitted the growth of magnetite.

Petrographic examination did not support the concept that the majority sample 600-630°C unblocking temperature of the characteristic remanence is due to maghemite. No unambiguous maghemite was observed. Since the most abundant opaque oxide observed petrographically was hematite in the form of martite and since unblocking temperatures for characteristic remanences higher than the 630°C ceiling do not exist, the most logical conclusion is that the main remanence carrier is a titanohematite (hemoilmenite). The microprobe analyses do indeed show that the majority of the opaque iron oxide grains are titanohematites. The microprobe analyses and petrographic observations of abundant martite suggests that the remanence carrier is a range of titanium-bearing hemo-ilmenites with sufficient titanium in the lattice to preclude Neel temperatures higher than 600-630°C. A Neel temperature of 600°C corresponds to a Ti-hematite with 14.5% Ti substituted for Fe in the hematite lattice [see *Butler*, 1992]; 630°C indicates ~11% substitution. Titanium-bearing hematites can only be formed by a high-temperature oxidation-exolution process because titanium cannot be incorporated into the lattices of iron oxides at low temperatures [*Buddington and Lindsley*, 1964].

Deposition of the lower Chinle Group was accompanied by abundant volcanic input; volcanic ash is widespread throughout the Petrified Forest Formation and its equivalents [*Stewart et al.*, 1972; 1986]. Much of the Petrified Forest claystones are devitrified tuffaceous material [*Schulz*, 1963]. The underlying Shinarump Formation has pebbles of rhyolitic composition, whereas the volcanic material in the Petrified Forest Formation is intermediate in composition, quartz latite or dacite [*Schulz*, 1963]. Intermediate igneous compositions would contain opaque oxides of both hematite and magnetite. *Stewart et al.* [1986] estimate the volume of volcanic debris supplied to the Petrified Forest Formation and equivalents was 70,000 km^3. The titanium content in the iron oxides of the Petrified Forest Formation, the rounded aspect of the grains, and the abundance of intermediate composition volcanic debris indicates that the dominant titanohematite remanence carrier, as well as the subordinate titanomagnetite carrier, must have been detrital, strongly suggesting that the Petrified Forest remanence is largely a depositional remanent magnetization.

4.2. Characteristic Directions and Paleopoles

Paleopoles were calculated from only samples that had three or more collinear measurements and possessed <7° of error (MAD values) associated with the fit of the line to the points in their demagnetization trends. These directions are displayed for the two formation members in Figures 7a and 7b.

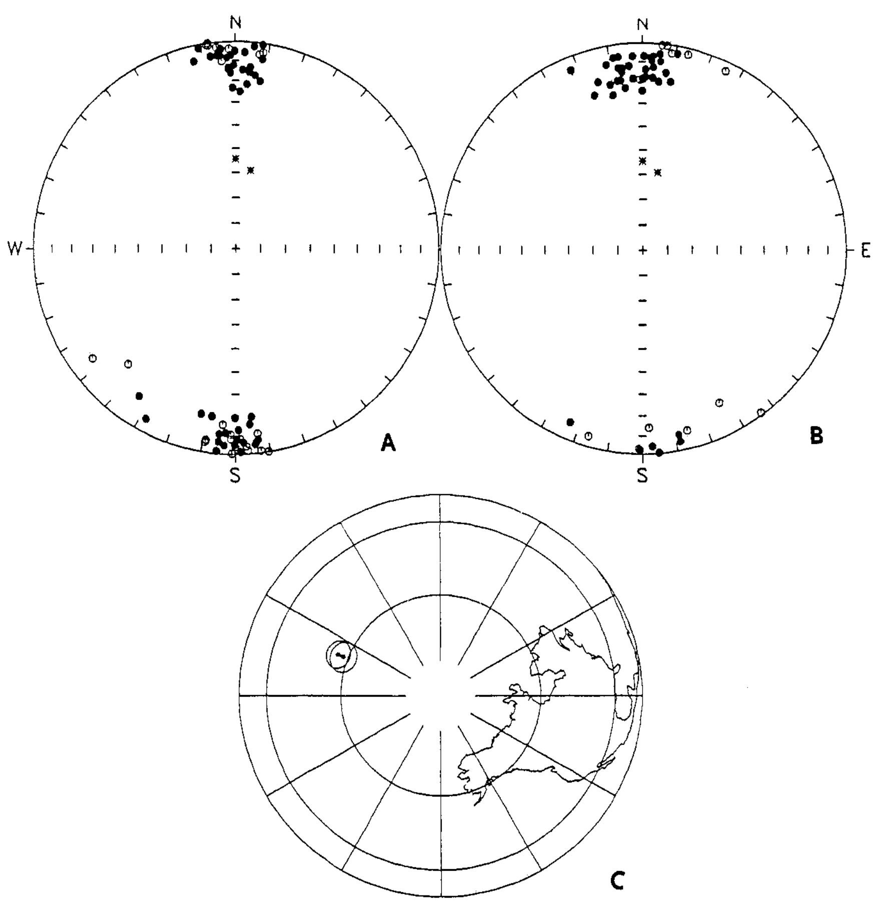

Figure 7. Characteristic magnetizations and paleopoles. (a) Lower Petrified Forest Formation (Blue Mesa Member); (b) Upper Petrified Forest Formation (Painted Desert Member); (c) paleopoles for the lower (L) and upper (U) Petrified Forest Formation and their alpha$_{95}$ confidence limits.

Table 1. Mean Directions and Pole Positions for the Petrified Forest Formation and Its Members

Computation	*D*	*I*	*N*	*R*	*K*	A_{95}	Long E	Lat N
Category I								
Blue Mesa Member	1.0	3.0	41	39.73	31.6	4.0	68.2	56.5
Painted Desert Member	0.8	6.8	24	23.52	47.5	4.3	68.5	58.4
Categories I and II								
Blue Mesa Member	1.4	2.4	70	107.26	29.4	2.5	67.5	56.2
Painted Desert Member	358.9	8.2	53	74.75	33.8	2.8	72.1	59.1
Test for antipolarity (category I samples)								
PF Formation normal	359.9	5.3	40	39.21	49.3	3.3	70.1	57.7
PF Formation reversed	182.5	-3.0	25	24.04	24.9	5.9	65.5	56.4
Ratio of kappas: 2.0; angle between means: 3.5°								
Petrified Forest Formation Paleopole:								
Category I	0.9	4.4	65	63.22	35.9	3.0	68.3	57.2
Categories I and II	0.1	4.9	123	118.50	27.3	2.5	69.8	57.5

Categories I and II represent data selections explained in the text. *D* and *I* are the mean declination and inclination; *N* is the number of samples; *R* is the length of the resultant vector; *K* represents the precision of the sample populations; A_{95} is the half-angle limits of 95% confidence; Long E and Lat N are the east longitude and north latitude of the paleopole locations; PF, Petrified Forest. Antipodal statistics computed from *McFadden and McElhinny* [1990].

Qualitative distinctions were made among these characteristic directions of Figure 7 on the basis of two criteria, the number of collinear demagnetization points and the amount of error (MAD values) associated with the line fit. As a result, samples were divided into two categories, I and II, in which category I samples are considered to have better defined remanences than those of category II. Category I consists of sample directions calculated from least 5 or more (up to 11) collinear points and errors/MAD values <5° and generally having values of 1-2°. Category II includes sample directions of four types: (1) samples with only 3-4 collinear points; (2) samples that had errors of 5-7°; (3) samples with low error and/or numerous points, but displaying linear behavior not directed to the origin; and (4) samples that displayed linear-to-the-origin behavior and low errors but whose resultant directions were a standard deviation or more away from those of category I. The latter behavior was included in category II because such samples with depositional inclination error, orientation errors, or a secondary remanence with stability comparable to the primary would be characterized by such a deviation.

Mean directions and pole positions were computed by Fisherian statistics (Table 1). Poles were calculated both from only category I samples ('I' in Table 1) and from a combination of categories I and II samples ('I&II' in Table 1). Separate paleopoles were calculated for the upper and lower members of the Petrified Forest Formation because of the age division between the two, with the lower, Blue Mesa Member being late Carnian, whereas the upper, Painted Desert Member is early Norian.

Despite the age difference, the two category I member paleopoles are statistically indistinguishable (Figure 7c). The two category I&II member poles also are statistically indistinguishable but are visibly offset slightly from one another (Table 1). This offset probably is the result of incomplete cleaning or other problems ascribed to category II samples but also may be the effect of an imbalance of polarity representation in the upper member population, there being a significantly greater proportion of normal polarity samples. Nevertheless, because the paleopoles from the two members are statistically identical, a single paleopole was calculated for the formation; again, mean formation poles were calculated from both I and I and II sample groups (Table 1).

Normal and reversed category I samples of the two members are antipodal; a positive reversal test [*McFadden and McEhlinny*, 1990] was obtained (Table 1). The angle between normal and reversed category I samples (n=65) was 3.5°; addition of the category II samples in an attempt to create a larger sample population (n=123) resulted in an angle of 10.2°, and probably reflects lack of removal of secondary magnetizations. The clearly antipodal nature of the category I samples and the excellent agreement between category I member poles make the category I pole preferable as the representative Petrified Forest Formation paleopole.

4.3. Magnetostratigraphy

The magnetic polarity of each sample was interpreted from the characteristic remanence. Figure 8 shows the magnetostratigraphic sequence of directions and sample lithologies, including stratal color. Polarity intervals were interpreted wherever three or more stratigraphic horizons yielded samples displaying the same polarity; indications of polarity change supported by samples from only two stratigraphic levels are shown as half bars in Figure 8. By this definition, eight polarity intervals are displayed by this lower three-fourths of the Petrified Forest Formation; if the polarity changes suggested by the samples from only two stratigraphic horizons are real, 15 total polarity intervals might have occurred during this time.

This polarity signature of the Petrified Forest Formation is well-defined with one exception. The upper part of the Blue Mesa Member between -21 and -53 m below the Sonsela Member (Figure 8) yielded a combination of fewer numbers of samples and samples more poorly magnetized than the rest of the formation. This interval is green claystone (part of unit 13 and unit 14 of *Lucas* [1993]) overlain by, and intertonguing with, blue/purple claystones (unit 15) and in turn overlain by cross-bedded friable white sandstone (unit 16). A significant portion of the strata was too fractured or friable to be sampled either by block sampling or core drilling. Commonly, weak remanence and inseparable multicomponent magnetizations characterize many of the samples obtained. As a consequence, this interval is represented by a much smaller number of samples and fewer still possessing well-defined remanence directions; only nine re-

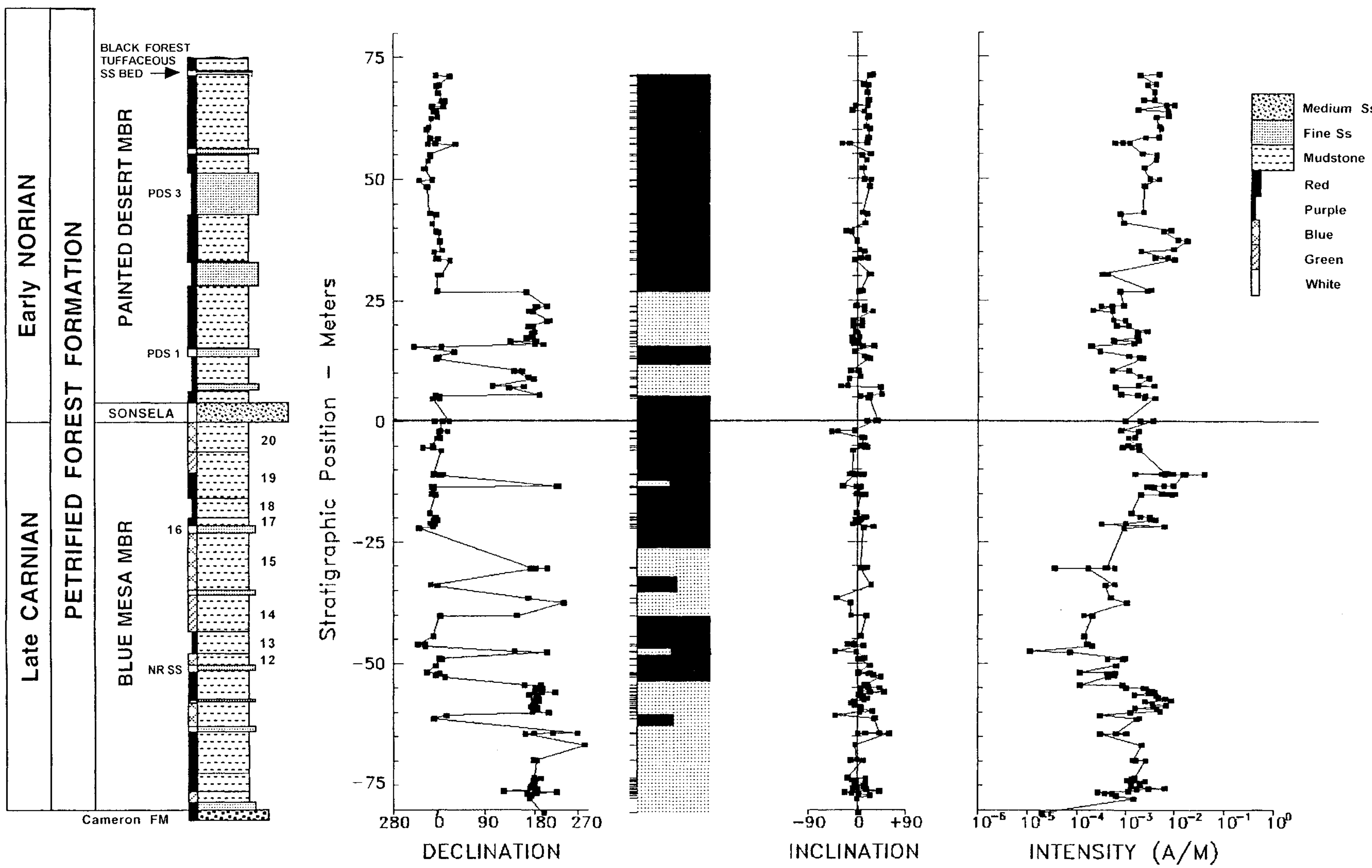

Figure 8. Magnetostratigraphic plots of the characteristic directions for both the lower and upper members of the Petrified Forest Formation. The middle, Sonsela Member lies at and above the line marking the Carnian-Norian boundary. The polarity interpretation column shows normal polarity as black and reversed polarity as a stippled pattern. Two widths to the polarity interpretation column signify levels of certainty of the polarity interpretation: full column indicates no uncertainty; 1/2 column width indicates potential short polarity intervals. Lithologic texture and color are indicated by patterns (see legend).

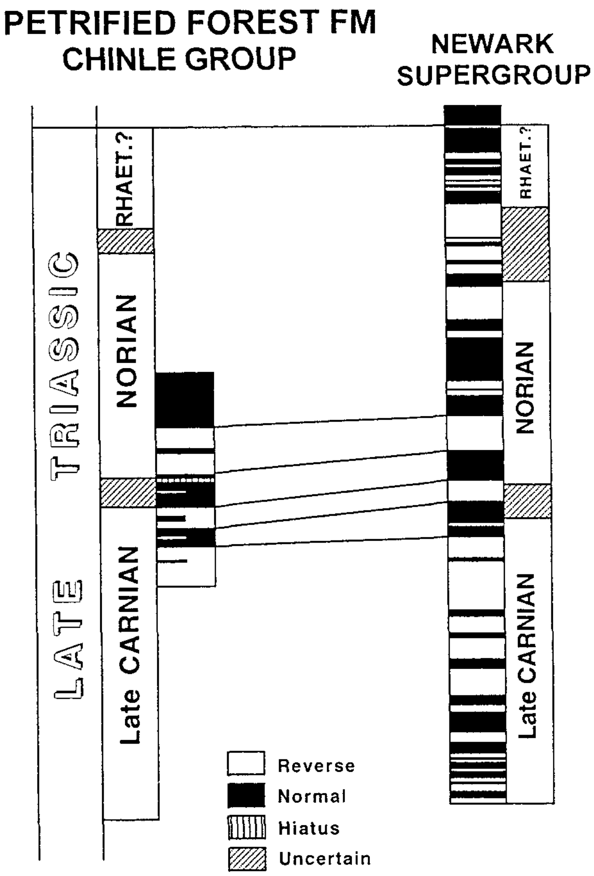

Figure 9. Comparison of the magnetostratigraphies of the coeval late Carnian-early Norian portions of the Chinle Group and Newark Supergroup.

liable samples represent the 17 meters between -23 and -40 m in Figure 8. This poorer magnetostratigraphic base means that geomagnetic field behavior might not be entirely accurate in this interval. It is both ironic and unfortunate that this interval which includes the best faunal data (unit 14) of the Petrified Forest Formation, the entire Chinle Group, and perhaps of the Late Triassic globally is not tied to good magnetostratigraphy.

The biochronology of the Chinle Group has been extended globally, including the Newark Supergroup; it is now clear that the Newark and Chinle strata span parts of exactly the same time interval [*Huber et al.*, 1993; *Lucas and Huber*, 1993; *Lucas et al.*, 1998]. The polarity sequence of the Petrified Forest Formation is coeval with that of the mid-portion of the Newark Supergroup. The two magnetostratigraphic records agree quite well (Figure 9). The fauna, flora, and their preservation within the Chinle Group are superior to those of the Newark Supergroup, hence the Chinle Group has more precision in biostratigraphic age. The greater abundance of tetrapod fauna makes the location of the Carnian/Norian boundary relatively well known in the Chinle Group, but less so in the Newark Supergroup. The full magnetostratigraphic and biostratigraphic correlation of the Chinle and Newark strata and its implications have been explored in a separate publication (M.B. Steiner, The Summerville Formation: A Middle Jurassic cratonic paleopole, rapid reversals, and the limited duration of the 'J5' unconformity, submitted to *Tectonics*, 1999); however, it is clear from magnetostratigraphic correlations between the two groups that the Chinle Group and Triassic Newark strata are precisely coeval.

It is of interest to note that the same polarity of magnetizations is present in the claystone intercalated with the lower sandstones of the Sonsela Member as in the immediately underlying claystones, because the Sonsela sandstones represent an abrupt change in depositional environment expressed by widespread deposition of coarser sedimentation onto a thick accumulation of fine-grained claystone. Moreover, this abrupt lithologic change coincides with the appearance of a new early Norian fauna above the Sonsela sandstones, in contrast to the late Carnian assemblage in strata beneath the Sonsela. Both lithologically and biostratigraphically, a break in the time record therefore is expected at the base of the Sonsela Member [*Lucas*, 1993; *Heckert and Lucas*, 1997]. The fact that Late Triassic normal polarity magnetization is recorded immediately below and above the presumed break may indicate a small or no time break, prevalence of a normal polarity geomagnetic field prevailed during a limited duration hiatus, or a large time break in which one or more polarity intervals are missing. A comparison of the entire Chinle magnetostratigraphic record with that of the Newark Supergroup (M.B., Steiner, Matched magnetostratigraphies in the Late Triassic Chinle and Newark Groups: Stratigraphic Completeness and the Late Triassic Magnetic Polarity Time Scale, submitted to *Geology*, 2000, Figure 2) supports the concept that the stratigraphic break represents little time.

4.4. The Late Triassic Paleopole

4.4.1. Slow APW. The Chinle Group is present on both the Colorado Plateau and the North American craton, and numerous paleopoles have been determined from both locations (formations yielding poles are indicated by asterisks in Figure 10a). It is well known that Colorado Plateau paleopoles are displaced with respect to counterparts from cratonic North America [*Steiner*, 1984; *Irving and Strong*, 1984; *Steiner*, 1986, *Bryan and Gordon*, 1986, *Kent and Witte*, 1993, *Molina-Garza et al.*, 1996]. Because the Chinle Group sampling sites of the present study are located on the Colorado Plateau, the Petrified Forest Formation paleopoles first are compared with other Colorado Plateau Chinle paleopoles (Table 2).

All Colorado Plateau Chinle paleopoles form a tight group (Figure 10b). Colorado Plateau paleopoles have been derived from three of the four tetrapod faunachrons in the Chinle Group; poles range in age from the Adamanian faunachron of the late Carnian to the Apachean faunachron of the Rhaetian ('Arizona-Utah' of Figure 10a and Table 2). The tight clustering eloquently confirms earlier conclusions of no, or very slow, apparent polar wander (APW) relative to the North American continent during the Late Triassic, a feature first observed from Newark Supergroup paleopoles [*Kent and Witte*, 1993; *Kent et al.*, 1995].

Paleopoles also have been obtained from the same three faunachrons of the Chinle formations residing on the North American craton ('Eastern New Mexico' of Figure 10a); these also cluster rather tightly (Figure 10c and Table 2). Thus the observation that little or no appreciable APW occurred relative to the North American plate during the Late Triassic now is indicated by three separate populations of Late Triassic paleopoles: those of the Plateau Chinle Group, the cratonic Chinle Group, and the Newark Supergroup. Mean Chinle Group paleopoles were derived separately for the Colorado Plateau and cratonic Chinle formations.

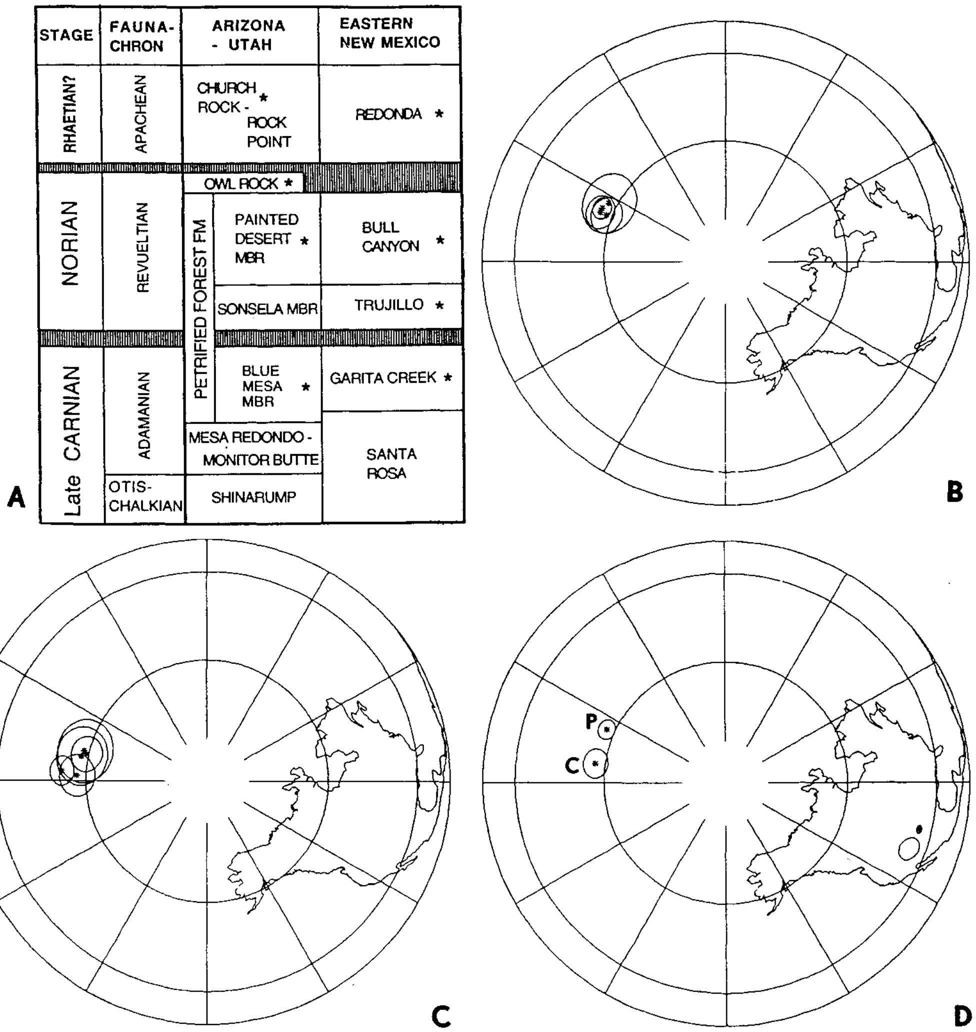

Figure 10. Comparison of Late Triassic Chinle Group pole positions. (a) Formations of the Chinle Group whose poles are plotted in Figures 10b and 10c; (b) Chinle Group paleopoles from the Colorado Plateau, including the poles of this study; (c) cratonic North American Chinle Group paleopoles; and (d) average Chinle poles for the NA craton (C) and Colorado Plateau (P), illustrating a separation of 9±3 °.

Statistically indistinguishable progressive eastward displacement of paleopoles relative to North America along the North American apparent polar wander path was observed in the Newark Supergroup [*Kent et al.*, 1995]. Kent et al. inferred the slight trend to indicate continuous very slow APW relative to North America during late Carnian through Rhaetian time in the Late Triassic. Such a trend was not observed in individual formation paleopoles from the Chinle Group but probably because all Chinle Group data are from surface exposures, whereas the Newark data were obtained from subsurface materials. Small unremoved recent secondary magnetizations in the Chinle strata probably mask this very small APW trend.

4.4.2. Colorado Plateau rotation. The relatively precise age equivalencies of respective Colorado Plateau and cratonic North American Chinle paleopoles and the very low rate of APW invite comparisons of mean faunachron paleopoles. The tetrapod biochronology of the Chinle Group allows age discrimination among formation paleopoles to the level of several million years; further, magnetostratigraphies associated with the Chinle paleopoles provide much more precise age comparisons, probably at the level of 100 to 250 kyr [*Steiner*, in review], between paleopoles of the Colorado Plateau and those of the craton. Comparison of cratonic late Carnian (Adamanian faunachron) paleopoles from the Garita Creek (Chinle Group) and Tecovas (Dockum Group) Formations [*Molina-Garza et al.*, 1995, 1996] to the Plateau Adamanian paleopole, the Blue Mesa Member of the Petrified Forest Formation of this study shows that they are separated by 8° (±4°). The succeeding Norian, Revueltian faunachron paleopoles on the Plateau are from the Painted Desert Member of the Petrified Forest Formation of this study and the superjacent Owl Rock Formation [*Bazard and Butler*, 1991]; these are separated from the cratonic Revueltian Bull Canyon Formation paleopole [*Bazard and Butler*, 1991] by 13° (±5 °). The paleopoles from the Rhaetian? (Apachean faunachron) Church Rock Formation of the Plateau [*Reeve*, 1975 (recalculated by *Gordon et al.* [1984]) and *Kent and Witte*, 1993] are identical and are separated from the cratonic Redonda Formation paleopoles [*Reeve and Helsley*, 1972 (recalculated by *Gordon et al.* [1984]) and *Molina-Garza et al.*, 1996] by 8° (±7°).

Thus Chinle faunachrons paleopoles indicate 8° to 13° differences between Colorado Plateau and craton Chinle paleopoles.

Table 2. Late Triassic Pole Positions

Age	Formation	Lat N	Long E	A_{95}	N	R	K	Reference
Chinle Plateau								
Apachean	Church Rock	63.3	57.5	7.3				1
Revueltian	Owl Rock	66.4	56.5	2.6				2
Revueltian	Upper Petrified Forest	68.5	58.4	4.3				this study
Adamanian	Lower Petrified Forest	68.2	56.5	4.0				this study
Mean Pole	Plateau Chinle	66.6	57.2	2.1	4	4.0	2635.2	
Chinle Craton								
Apachean	Redonda	79.0	58.0	7.0				3/4
Apachean	Redonda	76.5	58.5	7.6				5
Revueltian	Bull Canyon	87.8	57.4	5.0				2
Adamanian	Tecovas	77.7	59.3	4.2				6
Adamanian	Garita Creek	86.7	54.0	3.5				5
Mean Pole	Craton Chinle	81.7	57.5	3.3	5	4.99	531.2	
Newark Mean		98.3	56.2	2.6	7	6.99	539.2	

Symbols are as in Table 1. References are 1, *Kent and Witte* [1993]; 2, *Bazard and Butler* [1991]; 3, *Reeve and Helsley* [1972]; 4: *Gordon et al.* [1984]; 5, *Molina-Garza et al.* [1993]; 6, *Molina-Garza et al.* [1995].

The tight clustering of paleopoles from each tectonic block prompts computation of average Chinle poles for the craton and the Colorado Plateau (Table 2). It is clear (Figure 10d) that the Colorado Plateau and craton Chinle poles are not coincident, and are statistically quite distinct. The mean Plateau Chinle pole is displaced $9° \pm 3°$ clockwise from the mean cratonic Chinle pole. Chinle inclinations indicate that the Colorado Plateau resided in near-equatorial latitudes during Chinle deposition, the Plateau center being located at about 5°N paleolatitude, nearly 90° from the magnetic pole. A 9° counterclockwise rotation of the Plateau pole about an Euler pole at 34.6°N, 254.5°E [*Hamilton*, 1988] brings the two mean Chinle poles into excellent coincidence (Figure 11a).

The Chinle data thus indicate not only a clockwise rotation of the Colorado Plateau but a "large" clockwise rotation. The term "large" is contrasted to the smaller rotations of 5° or less advocated by *Bryan and Gordon* [1986, 1990], *May and Butler* [1986], and *Molina-Garza et al.* [1993, 1995, 1996]. The number of Chinle paleopole determinations involved in the present analysis is quite large, resulting in a very small statistical uncertainty regarding the location of Chinle mean poles. Chinle Group paleopoles therefore unequivocally demonstrate that the Colorado Plateau has experienced a larger rotation that conventionally believed.

The importance of accurately assessing the amount of rotation relative to the craton which the Colorado Plateau has experienced lies in the fact that the Plateau possesses a vast amount of undeformed stratigraphy from which paleopoles can be derived for a very large time period; knowing the exact amount of rotation will allow corrected Plateau paleopoles to be representative of the North American craton. Evidence that the Colorado Plateau has rotated by a magnitude greater than 5° also exists in Early Triassic, Middle Jurassic, and Late Jurassic strata [*Steiner*, 1998]. Magnetostratigraphy and biostratigraphy have demonstrated the age equivalence, hence appropriateness, of comparisons among these paleopoles. Colorado Plateau and craton paleopoles from the Early Triassic Moenkopi and Red Peak (Chugwater) Formations, the Summerville Formation, and the Salt Wash and Brushy Basin (lower and upper) Members of the Morrison Formation are shown with the Late Triassic Chinle average paleopoles in Figure 11b.

Magnetostratigraphy shows that the Early Triassic Red Peak Formation represents the same time period as the lower three-fourths of the Moenkopi Formation [*Steiner et al.*, 1993]. Considering the reversal frequency of the Early Triassic, the Moenkopi and Red Peak paleopoles are of equivalent age to within 0.5 m.y. or better. Moreover, within-formation studies of the position of the Early Triassic paleopole in a sedimentary record that encompasses most of the Early Triassic demonstrate that the Early Triassic also was a time during which North America experienced no APW [*Helsley and Steiner*, 1974; *Purucker et al.*, 1980; *Steiner et al.*, 1993]. Since no APW existed, any error in the age match of the Moenkopi and Red Peak Formations has no effect on conclusions regarding the amount of rotation of the Colorado Plateau inferred from Early Triassic paleopoles.

Magnetostratigraphy also has demonstrated the age equivalency of the Morrison Formation on the Colorado Plateau with that in cratonic eastern New Mexico [*Steiner et al.*, 1994]. Two distinctly different paleopoles were observed in sequential stratigraphic sequence in the formation on both tectonic blocks. Two lithologically distinct members yield the two poles in each case, and their magnetostratigraphies demonstrate synchronous deposition of each member on the Plateau and craton, demonstrating the validity of this member level paleopole comparison. Likewise, the Summerville Formation immediately underlies the Morrison Formation on both the craton and the Plateau and has produced paleopoles for each tectonic block.

Hence five sets of paleopoles display comparable significant differences between Plateau poles and craton counterparts. All coeval paleopole comparisons, individually and collectively, indicate a relatively "large" clockwise rotation of the Colorado Plateau. The five sets of poles can be brought into statistical coincidence by a $10° \pm 3°$ rotation of Plateau poles around *Hamilton's* [1988] Euler pole of 34.6°N, 254.5°E (Figure 11c). The $10° \pm 3°$ difference between equivalent Colorado Plateau and cratonic paleopoles obtained in this analysis is similar, although somewhat smaller than earlier "large" estimates: the original 13° suggested by *Steiner* [1984], the $15° \pm 4°$ concluded by *Irving and Strong* [1984] on the basis of Permian paleopoles, the $11° \pm 4°$ subsequently advocated by *Steiner* on the basis of Early Triassic and other paleopoles [1986, 1988], and the $13.5° \pm 3.5°$ concluded by *Kent and Witte* [1993] from a comparison of Newark and Chinle paleopoles. However, as emphasized before [*Steiner*,

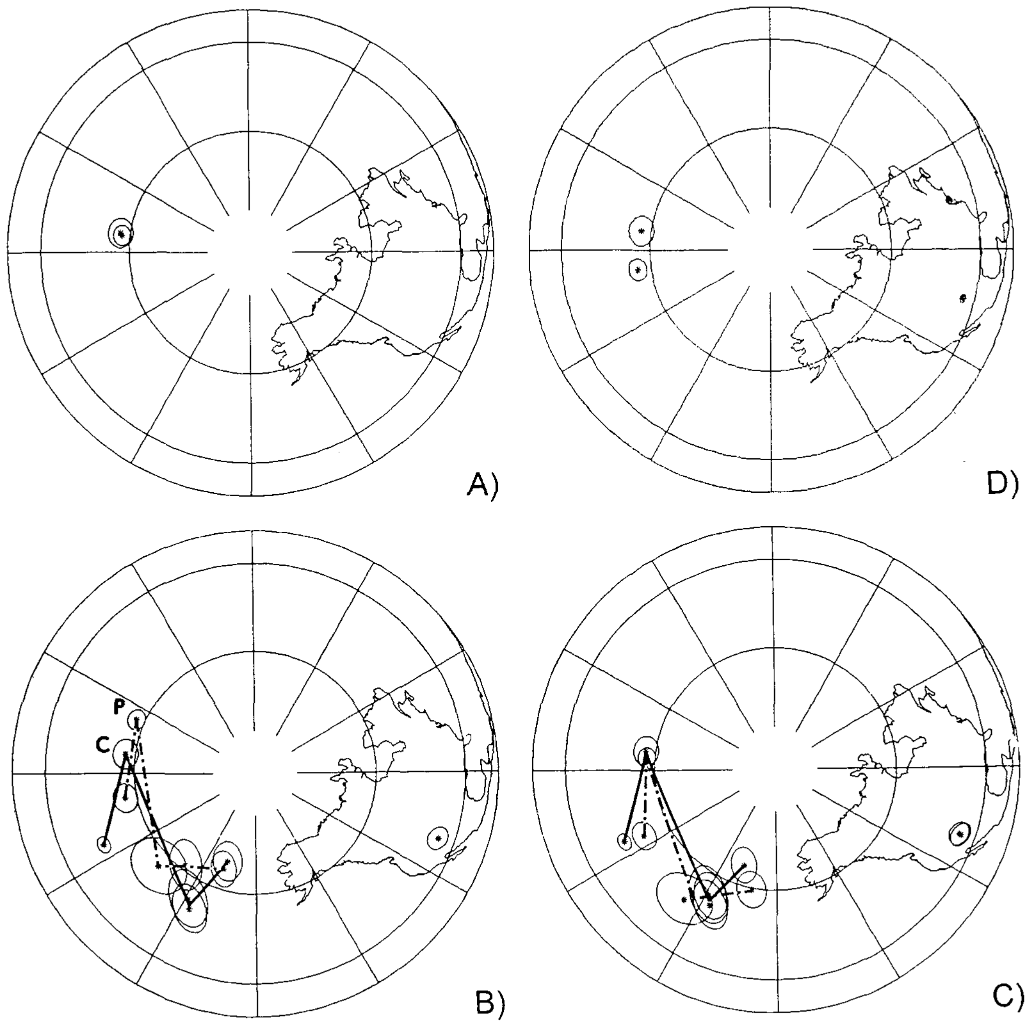

Figure 11. (a) Coincidence of the mean Colorado Plateau and cratonic Chinle paleopoles effected by a 9° counter-clockwise rotation of the Plateau pole (see text). (b) Comparison of paleopoles from coeval formations which are located on both the Colorado Plateau and the North American craton. The solid line connects cratonic paleopoles, illustrating North American apparent polar wander from this subset of data, whereas the dashed line traces the coeval Colorado Plateau APW path. The paleopoles displayed are the Early Triassic Moenkopi (Plateau) and Red Peak (craton) Formations, the Late Triassic Chinle Group, the Middle to Late Jurassic Summerville Formation, and the lower and upper parts of the Late Jurassic Morrison Formation [adapted from *Steiner* [1998]). (c) Counterclockwise rotation of Colorado Plateau paleopoles by 10$\pm$3 ° brings all Plateau poles into reasonable coincidence with corresponding craton poles (see text). (d) Comparison of the average Newark and cratonic Chinle paleopoles (N, Newark; P, Colorado Plateau Chinle; C, cratonic Chinle).

1998], even though a 10° $\pm$3° rotation of the Colorado Plateau describes the misfit of paleopoles from formations magnetized since the beginning of the Mesozoic with those of cratonic North America, the rotation was not a single event but consisted of at least two, and perhaps three, separate rotations.

4.4.3. Newark-Chinle pole discrepancy. In the course of this analysis, another, exceedingly surprising difference in paleopoles became evident. Even though biochronology and magnetostratigraphy demonstrate that the Chinle Group and the Newark Supergroup span precisely the same time interval [*Steiner*, in review], a statistically significant difference exists between their respective representations of the Late Triassic paleopole relative to North America: the Newark paleopole is displaced 9° counterclockwise from the cratonic Chinle paleopole (Figure 11d).

The comparison of Chinle and Newark paleopoles was achieved by calculating an average Newark Group paleopole (Table 2) from the seven Newark Basin paleopoles of *Kent et al.* [1995]. (A second Newark average pole also was calculated using these poles plus the lower, middle and upper Newark outcrop poles of *Witte and Kent* [1989] and *Witte et al.* [1991] and the Dan River pole [*Kent and Olsen*, 1997], with statistically indistinguishable results.)

Both the Chinle and Newark paleopoles are exceedingly well defined; an overabundance of data contributes to each. The magnetostratigraphic and biostratigraphic matches (Steiner, submitted manuscript, 2000) leave no doubt that these two groups were deposited at the same time, and at a time of little APW. The difference between their paleopoles indicates that relative motion has occurred between the Newark Basin of eastern North America and the western part of the North American craton. A clockwise rotation of the Newark Supergroup mean paleopole by 9° ±3° about various Euler poles located in the general Colorado-New Mexico-Texas region brings the mean Newark and cratonic Chinle paleopoles into good statistical coincidence. Of course, other poles of rotation are equally likely; one possibility is a vertical axis rotation of the Newark Basin.

5. Conclusions

The characteristic remanence of the Petrified Forest Formation is stable to thermal demagnetization temperatures of 600° to 630°C. Alternating field demagnetization and isothermal remanent magnetization (IRM) acquisition indicate the presence of a low-coercivity mineral (magnetite or maghemite) in several facies. IRM acquisition also demonstrates that both low- (magnetite or maghemite) and high- (hematite) coercivity minerals coexist in the red and purple facies. Reflected-light microscopy shows abundant martite, and high-temperature-exolved Ti-magnetite. Most Fe-oxide grains are rounded in nature. Microprobe analyses indicate that the dominant iron oxide phase is a titanohematite. The presence of titanohematite and titanomagnetite is consistent with the abundant intermediate-composition volcanic debris in the Petrified Forest Formation. Hence, the dominant carrier of characteristic magnetization in the Petrified Forest Formation is a detrital titanohematite, with contributions from magnetite in green and some portions of the red strata.

A minimum of eight polarity intervals of the geomagnetic field occurred in late Carnian through early Norian time. Paleomagnetic poles are identical between the late Carnian and the early Norian parts of the Petrified Forest Formation. Paleopoles of the Chinle Group indicate that North America experienced little or no apparent polar wander during the Late Triassic (late Carnian through Rhaetian). Coeval Chinle paleopoles of the Colorado Plateau and the cratonic North America require the Colorado Plateau to have experienced 9° ±3° of clockwise rotation since deposition of the Chinle Group. Combining five paleopole sets from Mesozoic formations deposited on the both Colorado Plateau and the North American craton indicates that the Plateau has rotated 10° ±3° since the Jurassic, although not necessarily all at the same time. Moreover, Chinle and Newark paleopoles, which share the same magnetostratigraphy and biostratigraphy, indicate that the Newark Basin has been displaced 9° ±3° counterclockwise relative to the North American craton.

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