JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 108, NO. B4, 2181, doi:10.1029/2002JB001909, 2003

Paleomagnetism and magnetostratigraphy of the lower Glen Canyon and upper Chinle Groups, Jurassic-Triassic of northern Arizona and northeast Utah

Roberto S. Molina-Garza
Centro de Geociencias, UNAM Campus Juriquilla, Querétaro, México

John W. Geissman
Department of Earth and Planetary Sciences, University of New Mexico, Albuquerque, New Mexico, USA

Spencer G. Lucas
New Mexico Museum of Natural History and Science, Albuquerque, New Mexico, USA

Received 1 April 2002; revised 7 October 2002; accepted 12 December 2002; published 2 April 2003.

[1] Twenty-eight selected sites (individual beds) in the Moenave Formation at the Echo Cliffs, northern Arizona, strata give a Hettangian paleomagnetic pole at 63.7°N, 59.7°E (dp = 2.6°, dm = 5.1°). The Wingate Sandstone and Rock Point Formation at Comb Ridge, southeast Utah, provide a Rhaetian paleopole at 57.4°N, 56.6°E (N = 16 sites; dp = 3.4, dm = 6.5). High unblocking temperatures (>600°C), high coercivity, and data analyses indicate that the characteristic magnetization is primarily a chemical remanence residing in hematite. The Hettangian and Rhaetian poles are statistically indistinguishable (at 95% confidence), they resemble existing data for the Glen Canyon Group, and they provide further validation to the J1 cusp of the North American apparent pole wander path (APWP). The red siltstone and upper members of the Chinle Group, on the south flank of the Uinta Mountains, northern Utah, define a Rhaetian pole at 51.6°N, 70.9°E (N = 20 sites; dp = 3.5°, dm = 6.9°). The Gartra and upper members of the Chinle Group in the north flank of the Uinta Mountains, give paleopoles at 52.0°N, 100.3°E (N = 6 sites; dp = 5.4°, dm = 10.5°) and 50.9°N, 50.1°E (N = 5 sites; dp = 8.8°, dm = 17.5°), respectively. These data indicate no significant rotation of the Uinta Mountains with respect to the craton. In total, data for the plateau and its bordering region of Cenozoic uplifts support estimates of small rotation of the plateau and provide evidence against the hypothesis of a Late Triassic standstill of the North American APWP. Our magnetostratigraphic results are consistent with lithographic and biostratigraphic data that place the Triassic-Jurassic boundary within the Dinosaur Canyon Member of the Moenave Formation, not at a regional hiatus. *INDEX TERMS:* 1520 Geomagnetism and Paleomagnetism: Magnetostratigraphy; 1525 Geomagnetism and Paleomagnetism: Paleomagnetism applied to tectonics (regional, global); 1527 Geomagnetism and Paleomagnetism: Paleomagnetism applied to geologic processes; *KEYWORDS:* paleomagnetism, Triassic, Jurassic, magnetostratigraphy, Colorado Plateau

Citation: Molina-Garza, R. S., J. W. Geissman, and S. G. Lucas, Paleomagnetism and magnetostratigraphy of the lower Glen Canyon and upper Chinle Groups, Jurassic-Triassic of northern Arizona and northeast Utah, *J. Geophys. Res.*, *108*(B4), 2181, doi:10.1029/2002JB001909, 2003.

1. Introduction

[2] Apparent polar wander paths (APWPs) require constant revision as new paleomagnetic, tectonic, or chronological data become available. For the Triassic-Jurassic sector of the North America APWP, a large proportion of poles defining the path are derived from strata of the Colorado Plateau (CP) and immediately surrounding regions. In previous compilations of the North American APWP [e.g., *Irving and Irving*, 1982; *Harrison and Lindh*, 1982; *Gordon et al.*, 1984; *Besse and Courtillot*, 1988; *Van der Voo*, 1990], the stratigraphic ages of these strata are rarely more precise than at the epoch level, yet better stratigraphic resolution is often possible, which may improve the calibration of the APWP. Similarly, ambiguity in the outline of the APWP can be reduced if paleomagnetic sampling is focused on parts of the stratigraphic record not studied before or those strata previously studied in insufficient detail. Ironically, efforts to reexamine lower Mesozoic sedimentary sequences in southwest North America [e.g., *Bazard and Butler*, 1991; *Steiner and Lucas*, 1993, 2000;

0148-0227/03/2002JB001909$09.00

Molina-Garza et al., 1991, 1995, 1998], mostly motivated by concern over data quality from earlier studies that employed blanket demagnetization, have raised more provocative questions than they have answered.

[3] One of several questions raised by recent studies concerns the rate of North America APW. A fast rate is suggested by the progression of Triassic poles across northern Asia (Figure 1a), but this hypothesis has been recently challenged [*Kent and Witte*, 1993; *Steiner and Lucas*, 2000]. A related question concerns the abruptness of the change in the direction of APW near the end of the Triassic, which has been defined as the J1 cusp [*May and Butler*, 1986]. A change in direction of polar wander is clearly apparent in the raw data and most interpolated versions of the path, but some workers attribute the abruptness to artifacts resulting from tectonic rotation of the CP, inadequate correlation of Triassic and Jurassic strata in this region and, possibly, an overall low reliability associated with the CP data [*Kent and Witte*, 1993; *Steiner and Lucas*, 2000].

[4] Which of the existing models for interpolating the APWP best portrays the North America Mesozoic data set is still debated. The succession of North American poles is reasonably well described by the paleomagnetic Euler pole (PEP) model (Figure 1b) [*Gordon et al.*, 1984], but a path averaged by combining groups of poles of similar age [*Van der Voo*, 1993] also describes the path sufficiently well (Figure 1c). Although other approaches exist, the paths illustrated show the problems associated with the J1 cusp. The change in plate motion in the interpolated path of *Van der Voo* [1993] is less dramatic than the PEP path. Moreover, the specific "reference" poles at the apex of the J1 cusp differ by $\sim 15°$. The ambiguity results from the manner in which the paleomagnetic poles are combined.

[5] The paleopole for the Wingate Sandstone of the Glen Canyon Group [*Reeve*, 1975] was used by *Gordon et al.* [1984] to anchor the J1 cusp in the North America APWP. This pole is of low reliability, as noted by *Van der Voo* [1990, 1993], and lies among a group of CP poles displaced westward with respect to coeval poles for sites on the craton interior. The existence and general location of the J1 cusp were later confirmed when data for the Hettangian Moenave Formation became available [*Ekstrand and Butler*, 1989].

[6] As mentioned above, CP data are displaced westward from coeval poles obtained in cratonic North America, which probably reflects small clockwise (tectonic) rotation of the plateau [*Gordon et al.*, 1984; *Steiner*, 1988; *Molina-Garza et al.*, 1998]. The magnitude of rotation of the CP remains, indeed, a distracting issue. Although those who have examined the structural and tectonic implications of CP rotation support a small (<4°) angle of rotation [e.g., *Bird*, 1998; *Cather*, 1999; *Wawrzyniec et al.*, 2002], the

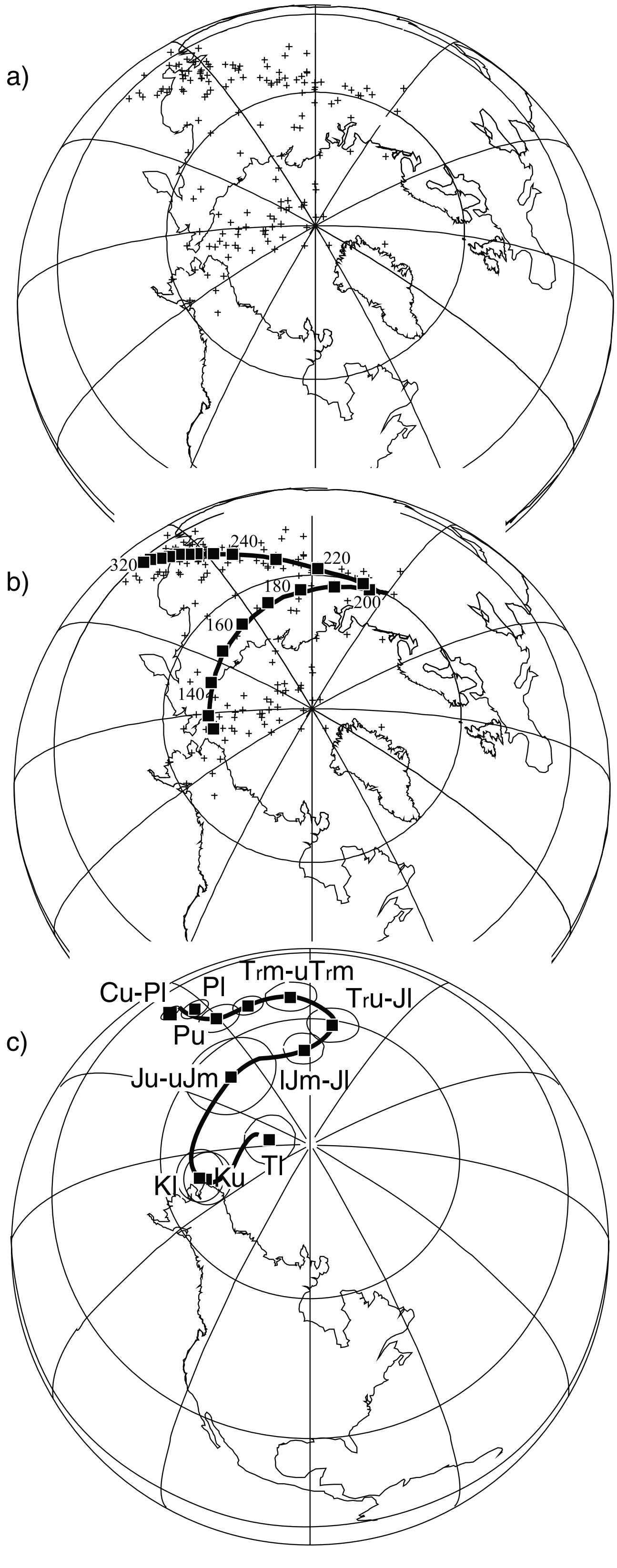

Figure 1. (opposite) (a) Late Paleozoic through Tertiary paleomagnetic data for North America. Poles from locations on the Colorado Plateau were restored for 5° of clockwise rotation of this region. (b) Paleomagnetic Euler poles of *Gordon et al.* [1984] superimposed on the raw paleomagnetic poles for the region. (c) Apparent polar wander path of *Van der Voo* [1993] based on average of poles falling within arbitrary age groups.

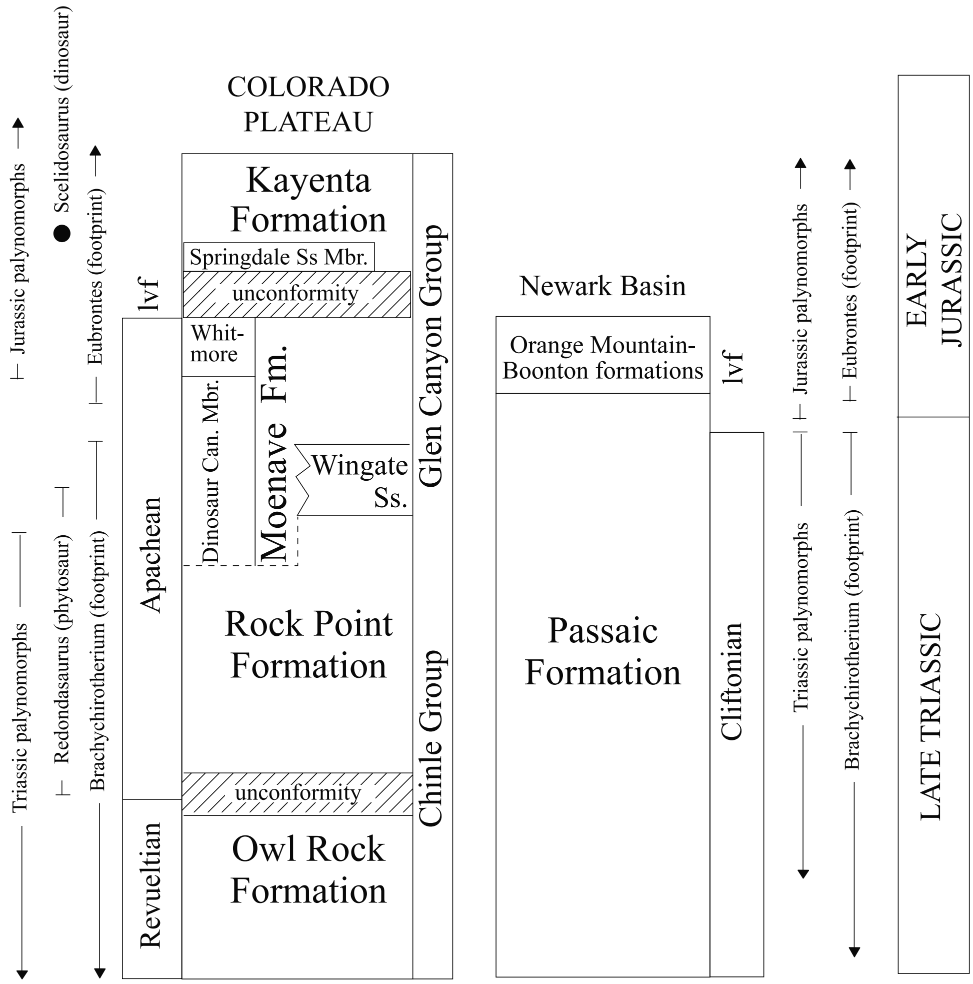

Figure 2. Correlation of Triassic-Jurassic strata on the Colorado Plateau with those in the Newark basin based on tetrapod biostratigraphy and palynostratigraphy. Lvf, land vertebrate faunachron. See text for discussion.

paleomagnetic community is still divided between those invoking small (<5°) [e.g., *Molina-Garza et al.*, 1998] and large (>10°) rotation [e.g., *Steiner and Lucas*, 2000]. One aspect that has been overlooked in this discussion is the fact that Laramide age deformation affected not only the margins [*Wawrzyniec et al.*, 2002] but also the interior of the CP. Therefore demonstrating the internal consistency of the plateau paleomagnetic database may shed new light into the rotation controversy.

[7] The rate of APW during Late Triassic and Early Jurassic time is not a trivial question, and claims of a fast rate of APW are independent of the actual amount of CP rotation [*Gordon et al.*, 1984; *Molina-Garza et al.*, 1995]. Data sets consistent with a rapid rate of Triassic-Jurassic APW include poles from the Upper Triassic Chinle Group and the Glen Canyon Group from sites on the CP and poles from age-equivalent strata on the High Plains of the craton interior. One data set that suggests a slow to negligible rate of APW during the Late Triassic was derived from the Newark succession in northeast North America [*Kent and Witte*, 1993]. We consider rotation of the plateau to be a less important issue than validation of the cusp as an abrupt feature in the North American APWP. Similarly important are absolute timing of the cusp and its duration. As noted by *Gordon et al.* [1984], the question of duration may relate to the time required for the North American plate to develop new plate boundary conditions. *May and Butler* [1986] also noted that a cusp of long duration may be viewed as an APW stillstand associated with an episode of intraplate deformation. Here, we present new paleomagnetic data for the Chinle and Glen Canyon Groups of the CP and surrounding region. These data allow us to examine the integrity and reliability of CP data as well as the correlation of lower Mesozoic strata in this region. We also discuss the implications for North America APW and Mesozoic magnetostratigraphy.

2. Geologic Setting and Stratigraphy

[8] The Late Triassic and Early Jurassic paleomagnetic record of western North America has been obtained from rocks of the Chinle and Glen Canyon Groups. Relevant litho- and biostratigraphic data are summarized in Figure 2. The Wingate Sandstone and the Moenave Formation are partially correlative, intertonguing, units at the base of the Glen Canyon Group [*Harshbarger et al.*, 1957; *Marzolf*, 1993].

[9] The Triassic-Jurassic boundary in southwest North America has been debated for decades. Uppermost Triassic

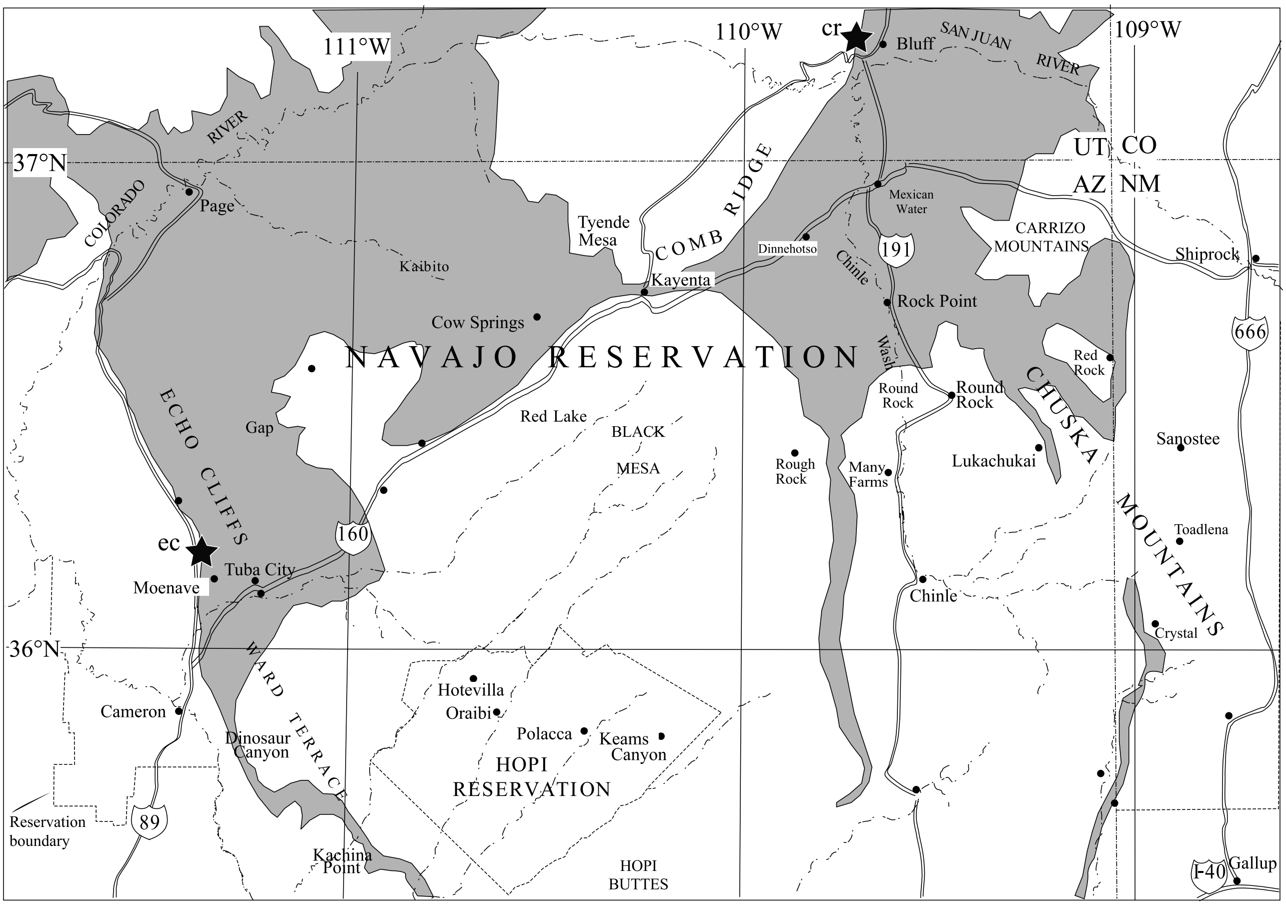

Figure 3. Location of sampling localities (stars) in northern Arizona and southeast Utah. The shaded region shows the outcrop distribution of Jurassic strata.

strata have been assigned to the Rock Point Formation of the Chinle Group. The Rock Point (also referred to as Church Rock Formation by some workers) underlies the Wingate Sandstone, but is absent throughout the Moenave outcrop area. The contact between the Chinle and Glen Canyon groups has been interpreted as either transitional [*Poole and Stewart*, 1964; *Harshbarger et al.*, 1957] or marked by a profound unconformity [*Pipiringos and O'Sullivan*, 1978]. Accordingly, the Triassic-Jurassic boundary has been variably placed above the Wingate Sandstone [*Harshbarger et al.*, 1957], at the Rock Point-Wingate contact, the so-called J-0 unconformity of *Pipiringos and O'Sullivan* [1978], below the Rock Point [*Lupe and Silberling*, 1985], and, based on the discovery of a phytosaur skull in lower Wingate strata, within the Wingate Sandstone [*Lucas et al.*, 1997].

[10] The pre-Early Sinemurian (Early Jurassic) age of the Moenave Formation and Wingate Sandstone is based on the recognition of *Scelidosaurus* in the Kayenta Formation; *Scelidosaurus* is known from lower Sinemurian marine strata of southwestern England [*Lucas*, 1996]. *Brachychirotherium* footprints and a skull of the phytosaur *Redondasaurus* indicate a Late Triassic age (Figure 2) for the lowermost Wingate Sandstone [*Lockley et al.*, 1992; *Lucas et al.*, 1997]. The Moenave Formation has traditionally included three members, the Dinosaur Canyon, Whitmore Point, and Springdale Sandstone members. The Dinosaur Canyon Member contains eolian strata considered by some to be a tongue of the Wingate Sandstone. The contact between the Springdale Sandstone Member and the underlying Whitmore Point Member is an erosional surface that has been interpreted by some as a major disconformity [*Marzolf*, 1993]. Lithologically, the Springdale is similar to overlying Kayenta sandstones; it is thus better considered part of the overlying Kayenta Formation [*Lucas and Heckert*, 2001]. The palynoflora of the Whitmore Point Member is *Corollina*-dominated and indicates an Early Jurassic age, but does not provide a stage-level age assignment [*Peterson and Pipiringos*, 1979; *Litwin*, 1986]. Tetrapod fossils from beds stratigraphically high in the Dinosaur Canyon Member, especially the crocodylomorph *Protosuchus* and the ichnotaxon *Eubrontes*, indicate that its middle to upper part is of Early Jurassic (Hettangian) age [*Lucas and Heckert*, 2001]. In summary, the Moenave Formation and Wingate Sandstone are bracketed as latest Triassic (Rhaetian) to earliest Jurassic (Hettangian). Clearly, the marine stages cannot be used in nonmarine strata in a meaningful way. Here, they are used in the sense applied to similar sequences such as the Newark Supergroup of eastern North America [*Olson et al.*, 2002].

[11] In northeast Utah, exposures of the Chinle Group and Glen Canyon equivalent strata are restricted to the flanks of the Uinta Mountains. Here, the contact between the Chinle Group and Glen Canyon strata is evidently conformable and

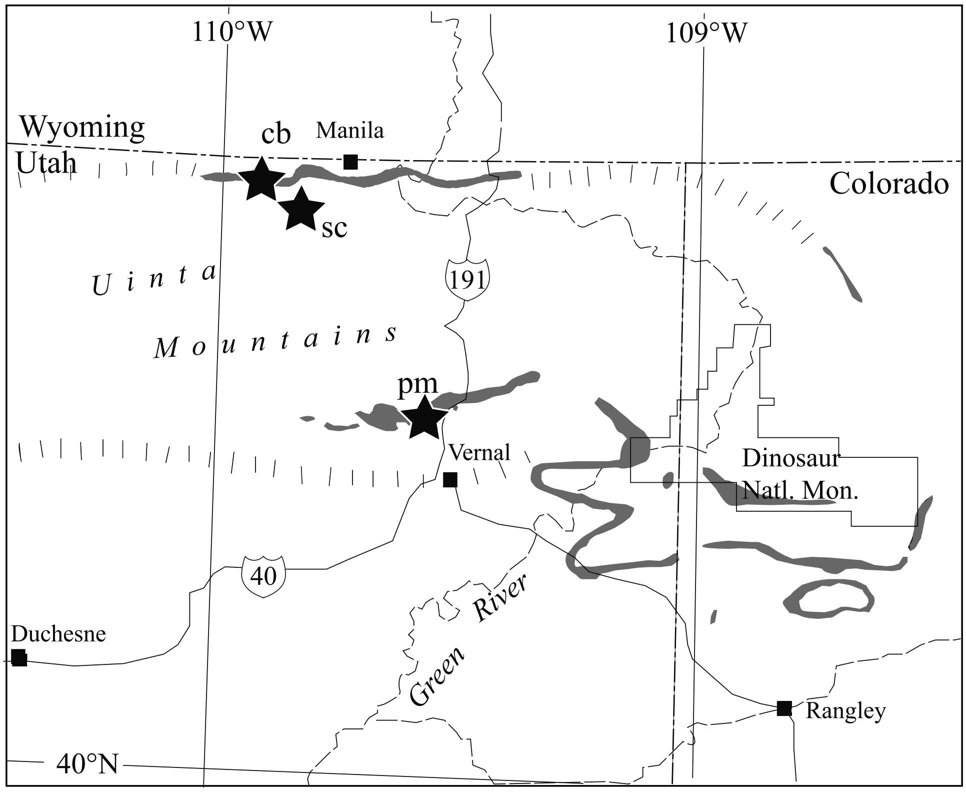

Figure 4. Location of sampling localities (stars) in northeast Utah. The shaded region shows the outcrop distribution of Upper Triassic strata, modified from *Poole and Stewart* [1964].

transitional [*Poole and Stewart*, 1964]. In this region, Upper Triassic strata are traditionally assigned, in ascending stratigraphic order, to the Gartra, ocher siltstone, sandstone-conglomerate, red siltstone, and upper members of Chinle Group strata [*Poole and Stewart*, 1964]. The red siltstone member has been correlated with the Church Rock Member of the Monument Valley and San Rafael Swell regions. *Poole and Stewart* [1964], and others, have suggested that Glen Canyon strata north of the Uinta Basin represent the lower part of this unit, specifically pre-Kayenta deposits.

3. Sampling and Methods

[12] For this study, we collected 4 to 10 oriented samples from a total of 39 paleomagnetic sites in Moenave and Owl Rock strata at the Echo Cliffs (EC, Figure 3), and 30 sites (one site equals one stratigraphic layer) in the uppermost Owl Rock, the entire Rock Point Formation, and the Wingate Sandstone at Comb Ridge (CR, Figure 3). Eight additional sites were collected from the Kayenta Formation at Comb Ridge, from a 40 m thick interval of conglomeratic sandstone that overlies Wingate eolianites. In both areas the rocks are well exposed, forming prominent cliffs. Samples were collected from both natural outcrops and road cuts using a portable drill. Samples were oriented using magnetic and sun compasses. Sites were obtained at irregular intervals mostly from well-indurated fine-grained sandstone and siltstone layers exposed in gently eastward dipping sections across the Chinle-Glen Canyon contact.

[13] At the Echo Cliffs, in north central Arizona (Figure 3), the Moenave Formation consists of a thin succession (30–115 m thick) of nonmarine, hematite-cemented, siliciclastic rocks, deposited in predominantly wet, eolian environments. The Whitmore Point Member pinches out a few miles east of Kanab, northeast of the studied section, into strata of the upper part of the Dinosaur Canyon Member. The Echo Cliffs lie on the eastern flank of the Kaibab uplift. This uplift, like others across the CP, may represent the early stages of Laramide foreland basement-cored uplift development [*Bump et al.*, 2000].

[14] At Comb Ridge, in northwest Arizona and southeast Utah (Figure 3), the Wingate Sandstone overlies Rock Point strata in a transitional contact. The Wingate consists of ~150 m of thick, massive, coarse to fine-grained eolian sandstone beds. Rock Point siltstone and sandstone beds (~50 m) overly the Owl Rock Formation of the Chinle Group at what appears to be a profound disconformity. The Comb Ridge monocline, a latest Cretaceous/early Tertiary Laramide style structure, lies on the steep eastern margin of the Monument Uplift of the CP.

[15] We also sampled uppermost Triassic strata at a total of 22 sites on the south dipping southern flank of the Uinta Mountains. The bulk of this collection is from a locality north of Vernal, from outcrops east of Highway 191 (Figure 4). Our sampling is limited to the red siltstone and upper members, which *Lucas* [1993] assigned to the Rock Point Formation. The stratigraphically highest strata in the Chinle Group consist of cross-bedded eolianites interbedded with pale red siltstone. A total of 12 additional sites were collected on the north dipping northern flank of the Uinta Mountain uplift. Seven sites were obtained in beds of the late Carnian Gartra Member at Connar Basin, ~10 km southwest of Manila, and

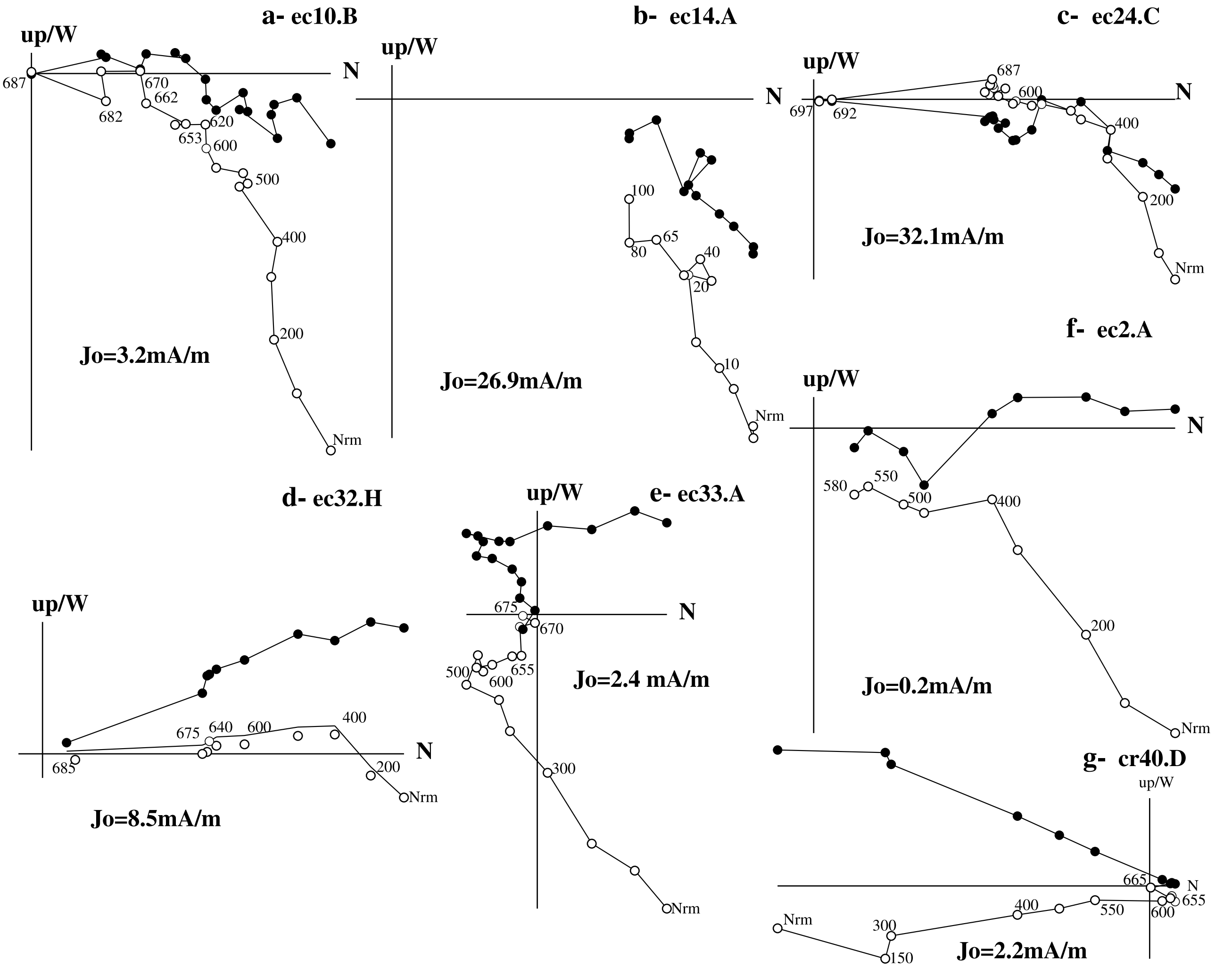

Figure 5. Orthogonal demagnetization diagrams for rocks in strata collected in the Moenave Formation at (a–e) the Echo Cliffs and (f–g) the Owl Rock Formation. Open (solid) symbols are projections on the vertical (horizontal) plane. All samples were thermally demagnetized samples except Figure 5b. Temperatures are in °C, and inductions are in mT.

five sites were collected at Sheep Creek Canyon from beds directly below the Chinle-Glen Canyon contact, south of Manila. The Uinta Mountain uplift is an east-west trending Laramide uplift that separates distinct tectonic domains. To the south lies the ‘‘undeformed’’ CP, and to the north lie a series of diverse-trending structural arches of the Rocky Mountain foreland [*Gregson and Erslev*, 1997]. The northern margin of the CP is typically placed along the southern front of the Uinta arch.

[16] In the laboratory, one or more specimens from each sample were prepared for NRM (natural remanent magnetization) measurements, which were carried out using a 2-G Enterprises 760R cryogenic magnetometer at the University of New Mexico. All samples were subjected to stepwise thermal or alternating field (AF) demagnetization. Chemical demagnetization was attempted in a small set of specimens but was relatively unsuccessful. Vector components to the remanent magnetization were identified from visual inspection of demagnetization diagrams [*Zijderveld*, 1967], and directions were calculated using principal component analysis (PCA) [*Kirschvink*, 1980]. Site means and the overall formation means were calculated using Fisher statistics [*Fisher*, 1953].

4. Demagnetization Results

4.1. Moenave Formation, Echo Cliffs

[17] The remanent magnetization of rocks of the Moenave Formation is of moderate intensity (10^{-3} to 10^{-4} A/m) and it normally consists of two vector components (Figure 5). A north directed and steep-positive magnetization of distributed laboratory unblocking temperatures overprints a north-directed and shallow characteristic magnetization (ChRM). The steep overprint is normally removed after heating to 450°C, but occasionally temperatures above 665°C are required to completely eliminate it. That is the case for sites in the lower Dinosaur Canyon Member (Figure 5a), and instability above this temperature leads to a high sample rejection rate (Table 1). A steep magnetization is also partly removed with alternating fields, indicating it resides in a cubic phase such as magnetite or maghemite (Figure 5b). This magnetization is interpreted as a secondary overprint of

Table 1. Paleomagnetic Data and Statistical Parameters for Site in the Echo Cliffs[a]

Site	$n\backslash n_d$	D, deg	I, deg	k	α_{95}	Tilt Correction
Moenave Formation (36.1°N, 111.4°W)						
Dinosaur Canyon Member						
ec6	6\6	357.1	10.3	327.3	3.7	
ec7	7\7	6.8	8.3	288.8	3.6	
ec8	6\7	1.7	12	53.5	9.2	
ec9	3\6	13.3	6.3	98	12.5	
ec10[b]	2\7	0.2	9			
ec11	3\7	19.2	9.1	230.3	8.1	
ec12[b]	5\7	8.6	8.2	6.8	31.6	
ec13[b]	4\5	357.3	23	9.8	30.9	
ec14	5\7	13.2	22.9	27.2	14.9	
ec15	6\8	4	28.4	10.3	23.2	
ec16	6\8	356.4	5.9	269.6	4.1	
ec17	6\8	345.8	39.3	26.3	13.3	
ec18	6\6	0	6.6	16.1	17.2	
ec19	4\4	353.7	3.8	16.7	23.2	
ec20	4\7	353.1	7.5	14.9	24.6	
ec21	6\6	350.2	−0.9	54.9	9.1	
ec22	6\6	2	10.5	266	4.1	
ec23	6\6	352.4	14.8	146.9	5.5	
ec24	7\7	357.3	4.6	37.8	9.9	
ec25	6\6	347.1	29.5	40.6	10.6	
ec26	6\6	350.4	11.6	39.5	10.8	
ec27	5\5	347.9	13.5	20.2	17.4	
ec28[b]	5\6	1.8	21.7	7	31	
ec29[b]	4\6	24.5	34.8	8.2	34.1	
ec30	6\6	341.5	5.7	20.3	15.2	
ec31	7\8	345.3	2	35.2	10.3	
ec32	7\7	335.8	−1.4	492.6	2.7	
ec33[b]	4\5	240.4	15.5	88.2	9.8	
ec34	4\5	7.5	20.6	73.7	10.8	
Springdale Sandstone						
ec35	7\7	347.3	2	81.8	6.7	
ec36	5\6	10.1	24	21.8	16.7	
ec37	6\8	9.3	25.3	45.1	8.3	
ec38	6\6	359.9	21.2	245.5	4.3	
ec39	3\3	2.2	18.3	321.3	6.9	
Overall mean	34	0.2	13.4	19.7	5.7	7.0, 18.6
Selected sites	28	357.4	13.1	31.0	5.0	4.3, 19.5
Sites ec6-11	6	8.6	12.0	58.6	8.8	14.2, 14.0
Sites ec15-32	16	351.4	11.3	34.6	6.4	357.8, 20.8
Sites ec34-39	6	2.5	18.8	47.3	9.8	11.6, 22.6

[a]Here $n\backslash n_d$ is the number of sampled usedνmber of samples demagnetized; D and I are the declination and inclination, respectively; k is Fisher's precision parameter; α_{95} is the radius of the confidence interval around the mean. The selected site mean yields a paleopole at 63.7°N–59.7°E.

[b]Indicates sites excluded from the final calculations, as explained in the text.

recent origin with a mean of D = 13°, I = 53° (n = 153 samples).

[18] The ChRM unblocks over a narrow range of temperatures, between about 600 and 690°C, and it is north-directed and shallow. The ChRM exhibits high coercivities (>100 mT) and most likely resides in hematite. Chemical demagnetization treatment is relatively unsuccessful and apparently removes a composite magnetization of intermediate inclination. One site in the upper Dinosaur Canyon Member (ec33) yields a southwest directed shallow ChRM, interpreted to record reverse polarity (Figure 5e). We note that in a previous study of the Moenave Formation in the Vermilion Cliffs of northwestern Arizona, *Ekstrand and Butler* [1989] found four sites carrying reverse polarity magnetizations in the uppermost 2 m of the Whitmore Point Member, a position equivalent to the upper Dinosaur Canyon Member.

[19] In most cases, the ChRM is well defined by linear segments in demagnetization diagrams that yield low MAD (maximum angular deviation) values. Nearly 60% of the 195 samples demagnetized yield MAD values <6°; 13 samples with MAD values >12° were excluded from site mean calculations. Seventeen additional samples did not yield useful results. Most site means are defined by five or more samples. Within site dispersion is highly variable; Fisher's precision parameter k ranges between 492 and 6.8 but it is <20 in eight sites. Four sites for which k is less than 10 and the confidence interval (α_{95}) is greater than 15° were excluded from the final calculations; site ec10 defined by only two samples was also excluded. Rocks of the Moenave Formation at the Echo Cliffs yield a tilt-corrected mean of D = 4.3°, I = 19.5° (N = 28 sites, k = 31.0, α_{95} = 5.0°; Table 1).

4.2. Owl Rock Formation, Comb Ridge and Echo Cliffs

[20] In the upper Owl Rock Formation, strata interpreted as silicified paleosol horizons are characterized by low NRM intensities (<~0.2 mA/m). Their magnetization is multivectorial, with a north directed and steep magnetiza-

tion overprinting a ChRM of shallow inclination (Figure 5f). The overprint is normally removed by heating to 400°C. In most of the sites from the Echo Cliffs, the overprint dominates the NRM (Figure 5f). Because of strong overprinting, the ChRM is poorly defined in sites ec1 and ec4; they were excluded from final calculations. The ChRM is of distributed unblocking temperature and is of both normal and reverse polarity. Data for the Comb Ridge and Echo Cliffs localities combined yield an ill-defined mean (tilt corrected) of $D = 11.8°$, $I = 19.4°$ ($N = 6$ sites, $k = 18$, $\alpha_{95} = 16.2°$). Besides being an imprecise result, we argue that the ChRM isolated in Owl Rock strata is unlikely to be an accurate estimate of the paleofield direction near the time of deposition. First, site mean inclinations are strongly bimodal; three sites yield normal polarity magnetizations of moderate inclination (~30°), and two sites of reverse polarity and one of normal polarity yield shallow inclinations (~5°; Figure 6). This results in a negative reversal test at the 95% confidence level. The moderate inclination magnetization of normal polarity is reminiscent of the magnetization termed "micrite component" by *Bazard and Butler* [1991] in a previous study of the Owl Rock Formation.

4.3. Rock Point Formation, Comb Ridge

[21] The ChRM of hematitic sandstone and siltstone beds of the Rock Point Formation at Comb Ridge is of high but distributed unblocking temperatures that range between ~300 and 670°C (Figure 7a). The NRM is overprinted by a steep north directed magnetization. The overprint is typically small, but occasionally, it is well developed and heating to ~600°C is necessary to completely unblock it (Figure 7b). AF demagnetization partially removes the steep overprint (Figure 7c), suggesting that part of it resides in magnetite or maghemite and part of it resides in hematite. The ChRM is generally south directed and of shallow inclination, except for a thin interval where, though poorly defined, a north directed and shallow magnetization is evident (Figure 7d). This thin interval of normal polarity in the Rock Point is identified in two sections at equivalent stratigraphic levels (cr49 and cr51), but its record is complex. A well-developed steep overprint dominates the remanence of these sites, and at both sites within-site dispersion is high. In at least one sample from each site, the ChRM is south directed and shallow. Thus both polarities may be recorded in this layer. We interpret these observations to indicate that the feature of normal polarity is of relatively short duration with respect to the time of remanence lock-in.

[22] The ChRM of the Rock Point Formation is well defined in 56 of 68 samples demagnetized. The quality of the data is evident in the fact that MAD values are <6° in ~50% of the 68 samples demagnetized. Fourteen samples were excluded because MAD values are greater than 12°. Despite of the low MAD values, a site mean is well-defined in only 6 of 11 sites. Two site means are defined by two samples and were excluded from the final calculations; 3 sites were also excluded because $k < 10$ and $\alpha_{95} > 15$. Six accepted sites yield a tilt-corrected mean of $D = 195.5°$, $I = -18.6°$ ($k = 52.7$, $\alpha_{95} = 9.3°$; Table 2).

4.4. Wingate Sandstone, Comb Ridge

[23] The NRM of Wingate strata is relatively straightforward. NRMs are of moderate intensity, between 0.5 and 5×10^{-3} A/m. The ChRM is a dual polarity magnetization of high discrete unblocking temperature (>600°C). The ChRM is overprinted by a prominent north-directed steep positive magnetization of high coercivity but low and distributed unblocking temperatures (Figures 7e and 7f). Seventy of 88 samples demagnetized were used in the final calculations. The ChRM is well defined (MAD <6°) in ~60% of the 88 samples demagnetized. Generally, AF treatment brings only a small to negligible reduction in magnetization, suggesting that magnetite is not an important remanence carrier in Wingate strata. Five samples were excluded because MAD values are >12°. Site means are well defined in 11 of 16 sites. Five sites (cr58–60, cr103, and cr125) were excluded from final calculations following the same criteria as in other localities. The polarity of these beds is, however, unambiguous. The tilt corrected mean of sites obtained in Wingate strata at Comb Ridge is $D = 182.9°$, $I = -5.9°$ ($N = 11$ sites, $k = 46.5$, $\alpha_{95} = 6.8°$). The angle between reverse and normal site means is 6.4°, a reversal test is positive, but the angle required to reject the common mean hypothesis is 14° [*McFadden and McElhinny*, 1990].

4.5. Kayenta Formation, Comb Ridge

[24] Demagnetization data for samples of the Kayenta Formation are of considerably lesser quality than other formations studied (Figures 7i and 7j). These coarse-grained sandstones do not respond well to thermal, AF, or chemical demagnetization. Thermal demagnetization to ~400°C removes a steep magnetization. Above 400°C, directions decay toward the origin, but decay is erratic due in part to the weak intensities of the samples. Demagnetization trajectories suggest that the ChRM remaining is shallow and north (or south) directed. The direction of the ChRM cannot be estimated with confidence, but the polarity at each site is unambiguous.

4.6. Red and Upper Members: Chinle Group, Southern Uinta Mountains

[25] The magnetization of Chinle strata in the southern flank of the Uinta Mountains is relatively simple. A small overprint of steep inclination, and a ChRM of high coercivity and high unblocking temperature characterize the red siltstone member (Figure 8a). A narrow range of laboratory unblocking temperatures, between 600 and 695°C, is typical of the ChRM at most of the sites. The few sites in pale orange sandstones, low in the red siltstone member, carry a larger overprint than brick-red siltstone and sandstone of the upper part of this unit (Figure 8b). The overprint is typically north directed and steep and is easily removed by heating to ~500°C.

[26] The NRM of samples collected in the upper member is also the sum of a steep (viscous?) overprint, with maximum laboratory unblocking temperatures of ~450° and a shallow, south-directed magnetization with a narrow range of unblocking temperatures between 600 and 690°C (Figure 8d). As in the red siltstone member, the high unblocking temperatures and high coercivity indicate that hematite is the sole carrier of the ChRM. The narrow range of laboratory unblocking temperatures suggests that contributions from specular hematite are important.

[27] The ChRM is south directed and of shallow positive and negative inclinations in all sites except site pm16, low

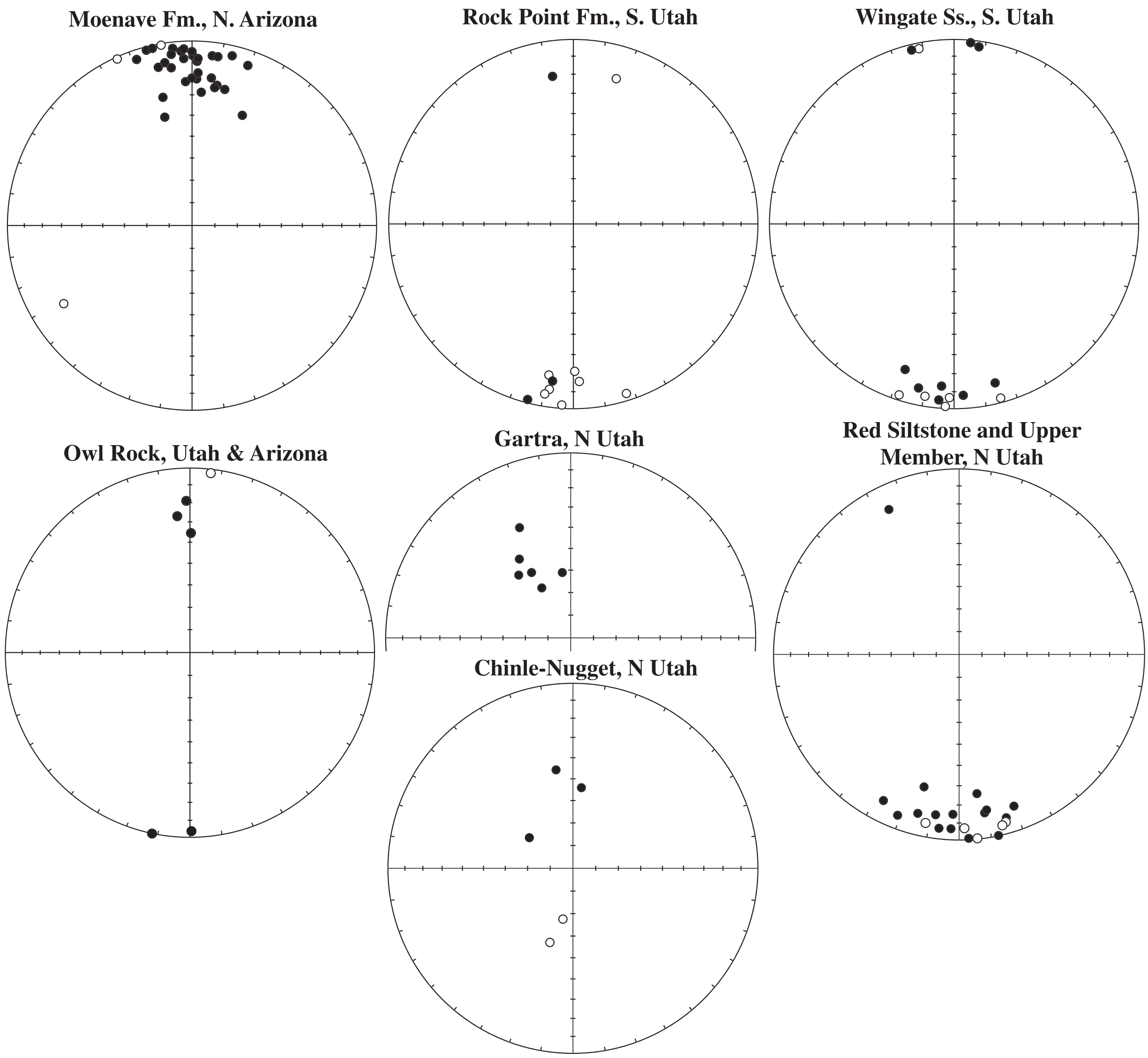

Figure 6. Equal-area stereographic projections of site-mean distribution for the localities studied. Open (solid) symbols are projections in the upper (lower) hemisphere.

in the upper member. The NRM at site pm16 consists of a steep overprint, a shallow north directed magnetization of intermediate coercivity (10–60 mT) and distributed laboratory-unblocking temperatures between 180 and 600°C, and the reverse polarity ChRM (Figure 8c). The shallow north directed magnetization is interpreted as a normal polarity ChRM but demagnetization data suggest that the north directed and shallow magnetization may be partially contaminated by a remanence of viscous origin.

[28] Data from the section collected along the southern flank of the Uinta Mountains are of the highest quality. The direction of the ChRM is well defined in nearly all samples demagnetized, with low MAD values ranging between 0.5 and 10° (averaging 5.5°). Only sites pm10 and pm16 were excluded from the overall mean. The site means are based on at least three samples per site (Table 3). Within-site dispersion is low, with k values ranging between 20.6 and 204.2. The in situ mean of 14 sites collected in the red siltstone member is of D = 178.8° and I = 8.5° (k = 21.4, α_{95} = 8.8°). Between-site dispersion is relatively high, but consistent with other studies of hematite bearing sandstones. Data quality is also high in the upper member, with MAD values averaging 4.3°. The upper member mean is of D = 180.9° and I = 7.0° (k = 25.0, α_{95} = 13.7°), and is based on data for six sites. The small number of sites explains the relatively large α_{95} value. Strata dip gently ~15° to the southeast; the corrected mean for the upper member is of D = 181.0° and I = −5.6° and the mean for the red siltstone member is D = 179.2° and I = −3.8°.

4.7. Gartra Formation and Upper Member: Nugget Sandstone Transition, Northern Uinta Mountains

[29] Sites collected in Connar Basin yield high-coercivity and high laboratory unblocking temperature magnetizations. The NRM of strata assigned to the Gartra Member is near univectorial, and is characterized by a narrow range of laboratory unblocking temperatures (Figure 8e). There is a hint of a small north-directed and steep overprint, but this is easily removed isolating a ChRM that is (in situ) northwest directed and of steep positive inclination. Site cb1, collected

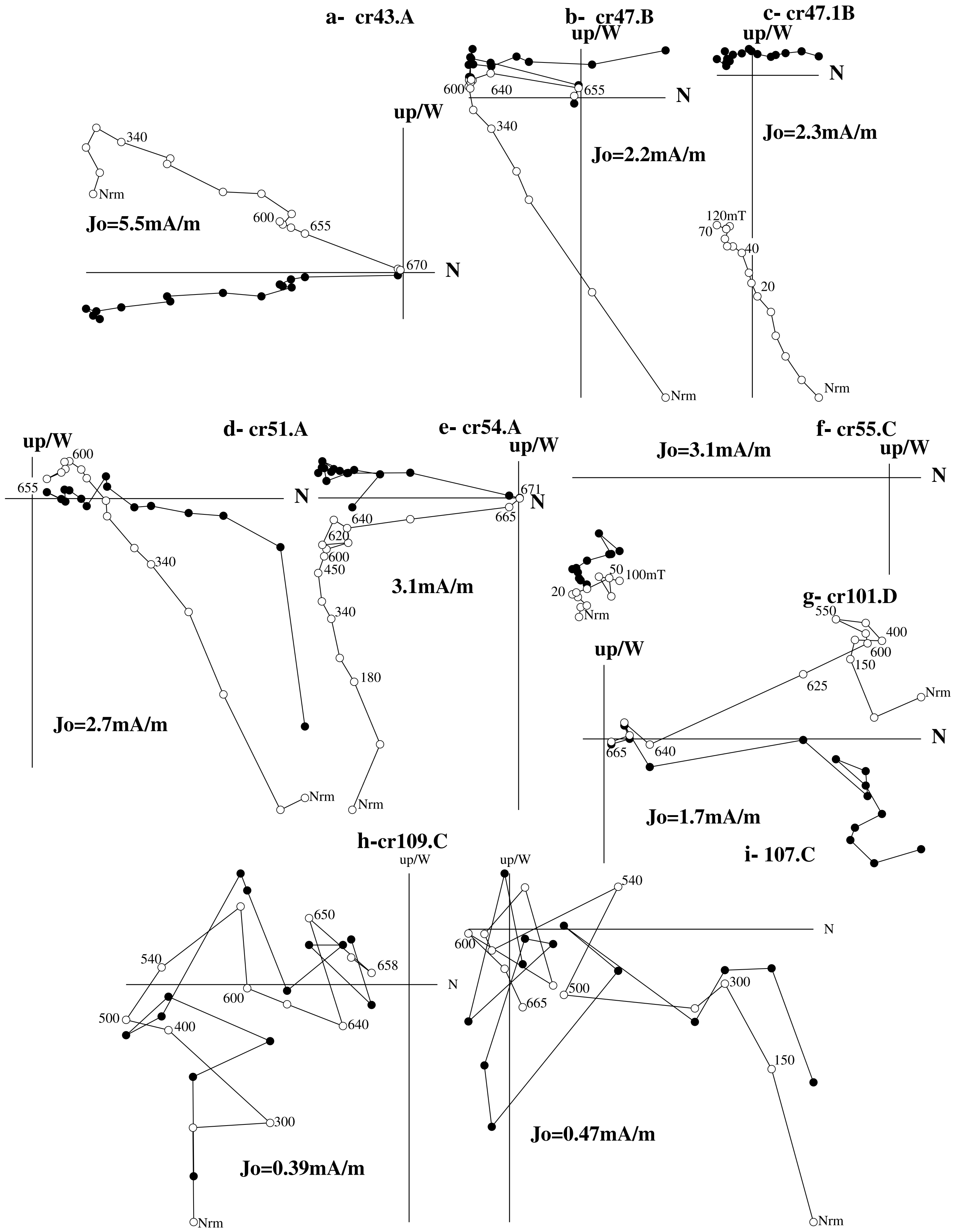

Figure 7. Orthogonal demagnetization diagrams for rocks in strata collected in the (a–d) Rock Point Formation, (e–g) Wingate Sandstone, and (h–i) Kayenta Formation at Comb Ridge. Open (solid) symbols are projections on the vertical (horizontal) plane. All samples were thermally demagnetized samples except Figures 6c and 6f. Other symbols as in Figure 5.

Table 2. Paleomagnetic Data and Statistical Parameters for Sites at Comb Ridge[a]

Site	$n \backslash n_d$	D, deg	I, deg	k	α_{95}	Tilt Correction
Wingate Sandstone						
(37.26°N–109.66°W)						
cr50	6\6	185	4.8	27.9	12.9	
cr54	7\7	192.2	10	77.7	6.9	
cr55	4\6	198.6	18	16.7	23.2	
cr56	7\7	165.5	12.1	101.1	6	
cr57	7\7	182.8	−1.1	78.1	6.9	
cr58[b]	3\6	181.6	−6.6	5.7	57.1	
cr59[b]	3\7	197.8	−3.1	8.3	45.9	
cr60[b]	4\7	165.1	−2.8	5.7	42.3	
cr61	6\6	184.4	13	31.4	12.1	
cr62	5\7	177	7.9	71.5	9.1	
cr63	3\3	189.6	−6.1	97.4	12.6	
cr101	4\5	8	−3.5	19.9	21.1	
cr102	4\5	346.2	−3.5	46.2	13.7	
cr105	3\5	5.2	−1.7	66.3	15.3	
cr125[b]	4\4	348.7	3.5	6.8	37.9	
Overall	15	181.8	3.5	36.7	6.4	183.6, −9.1
Selected	11	183.1	6.5	46.5	6.8	182.9, −5.9
Rock Point Formation						
cr42	7\7	189.3	−18.2	428.6	4.4	
cr43	7\7	179.5	−21.4	122.6	5.5	
cr44	5\7	177.8	−15.7	411.2	4.5	
cr45	7\7	183.7	−2.1	116.1	5.6	
cr46	6\6	194.6	2	69.7	8.1	
cr47	5\6	188.3	−10.5	13.4	26	
cr48[b]	3\3	162.6	−4.4			
cr49[b]	5\7	16.4	19.1	5.1	45.4	
cr51[b]	5\7	352	−20.5	5	38.2	
cr52[b]	4\5	187.6	15.2	7.4	49	
cr53[b]	2\6	189.6	−7.5			
Overall	11	183.8	−5.7	22.9	9.7	190.7, −15.4
Selected	6	185.6	−11.0	52.7	9.3	195.5, −18.6
Combined	26	182.7	−0.4	27.7	5.5	186.8, −11.8
Selected	17	184	0.1	32.5	6.4	187.4, −10.7
Owl Rock Formation						
ec2	4	5.8	−2.4	23.5	19.3	5.7, 1.9
ec3	3	0.7	35.5	58.2	16.3	18.9, 38.2
ec5	3	179.5	3.6	69.8	14.9	179.4, −3.3
cr40	4	191.8	0.2	30.7	16.8	193.7, −6.0
cr40n	3	354.6	26.8	244.8	7.9	23.3, 37.6
cr41	7	358	18.1	39.5	9.7	14, 28.7

[a]Here $n \backslash n_d$ is the number of sampled usedνmber of samples demagnetized; D and I are the declination and inclination, respectively; k is Fisher's precision parameter; α_{95} is the radius of the confidence interval of the mean. The combined mean of 17 selected sites gives a pole at 57.4°N–56.6°E.

[b]Indicates sites excluded from the final calculations, as explained in the text.

in coarse conglomeratic sandstones did not give any useful data, but a ChRM was isolated in the six remaining sites. All sites give remarkably well-grouped site means, with small α_{95} values (of 2.3° to 4.6°) and high precision parameters in excess of 300 (values not typical for red beds). Between-site dispersion, in turn, is moderately high but not unusual for this type of rock. The six sites give an in situ mean of D = 331.2° and I = 53.4° (k = 43.3, α_{95} = 10.3°). Strata dip 43° almost due north and give a corrected mean of D = 341.8° and I = 12.9°.

[30] Five sites collected in uppermost Triassic-Lower Jurassic strata immediately below the Chinle-Nugget Sandstone contact, in the north flank of the Uinta Mountains are characterized by two magnetization components. A north directed and steep positive inclination magnetization overprints a magnetization that is north to northwest (or south to south-southwest) directed and of steep positive (or negative) inclination. This dual-polarity magnetization is interpreted as the ChRM. The overall mean (N = 5 sites) is of D = 1.3° and I = 61.7°. Strata dip 54 to 69° to the north-northeast, and give a corrected mean of D = 12.6° and I = 6.8° (k = 20.0, α_{95} = 17.5°). Dip variation is insufficient to provide a statistically meaningful fold test.

4.8. Rock Magnetism

[31] Demagnetization results suggest that the ChRM of all of the rocks studied resides primarily, if not exclusively, in hematite. This is consistent with the high laboratory unblocking temperatures and high coercivities observed. Demagnetization data also suggest the presence of a low coercivity phase such as magnetite or maghemite; it carries a steep north directed overprint. Isothermal remanence acquisition curves (Figure 9a) indicate that inductions of 3 T are insufficient to reach saturation, and IRM intensities climb steadily with increasing inductions. Again, this indicates that a high coercivity phase dominates the magnetic mineralogy. AF demagnetization curves of the IRM induced with 3 T (IRM_{3T}) show that ~20% of the remanence is

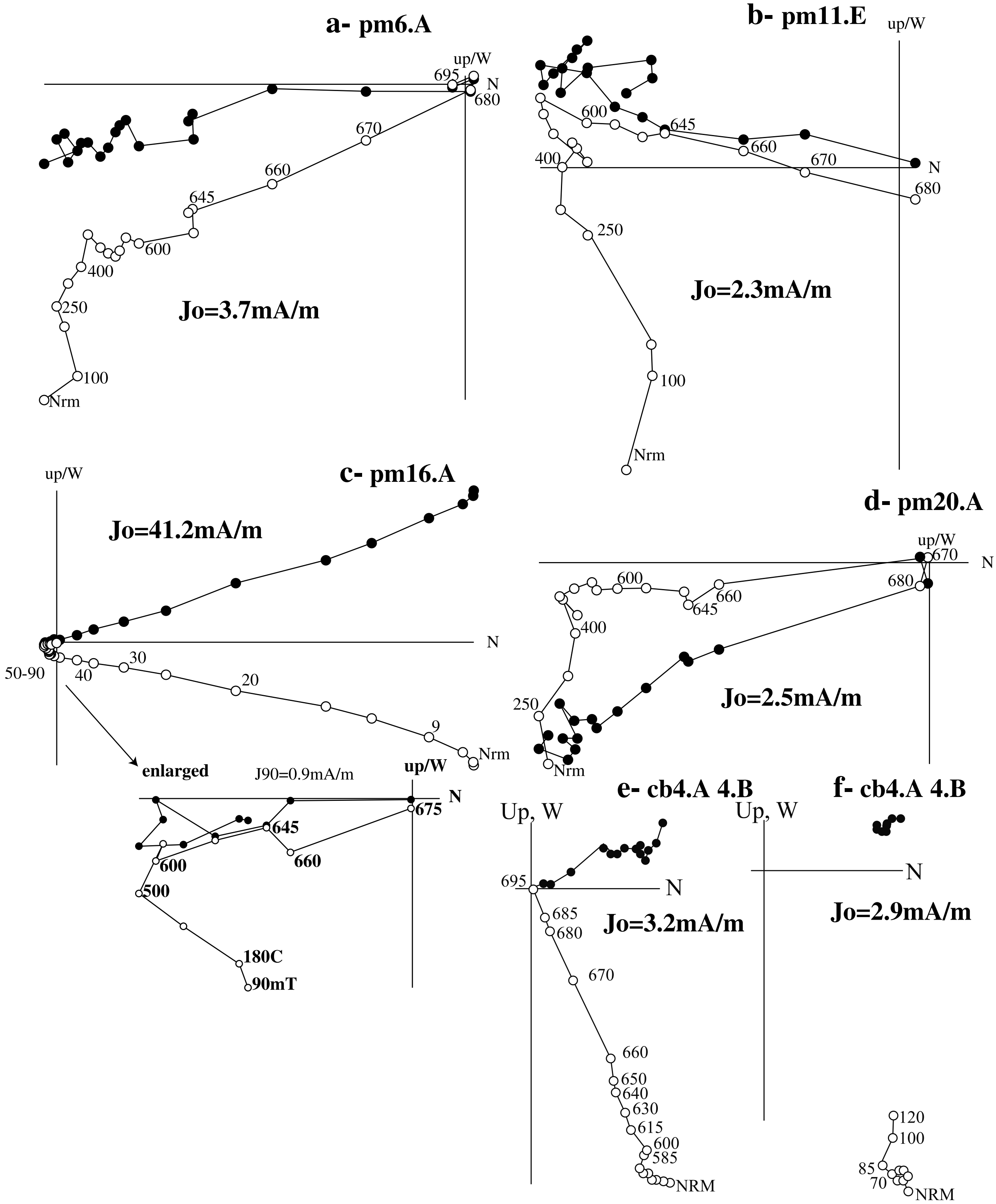

Figure 8. Orthogonal demagnetization diagrams for rocks in strata collected in the (a–b) Red Siltstone and (c–d) Upper members of the Chinle Formation in the southern flank of the Uinta Mountains, and (e) the Gartra Member in the northern flank of the range. Open (solid) symbols are projections on the vertical (horizontal) plane. All samples are thermally demagnetized samples except Figures 8c and 8f. Thermal demagnetization of Figure 8c after AF is shown in the detail enlarged. Other symbols as in Figure 5.

Table 3. Paleomagnetic Data and Statistical Parameters, Uinta Mountains[a]

Site	$n\backslash n_d$	D, deg	I, deg	k	α_{95}	Tilt Correction
Red siltstone member, north Utah (40.57°N, 109.58°W)						
pm1	4\4	159.8	14.2	71.8	10.9	
pm2	4\4	186.5	6.4	41.7	14.4	
pm3	4\4	174.2	−0.8	161.9	7.2	
pm4	4\4	165.6	−6	18.6	21.9	
pm5	4\4	167.5	0.4	38.2	15.1	
pm6	4\4	172.6	25.3	154.2	7.4	
pm7	4\4	169.9	16.2	56.6	12.3	
pm8	4\4	194.3	12.9	204.2	6.4	
pm9	4\4	188.2	14.2	54.8	12.5	
pm11	4\4	178.2	−7.1	20.6	20.7	
pm12	4\4	182.5	7	66.2	11.4	
pm13	4\4	207.3	12.5	48.7	13.3	
pm14	4\4	194.6	27.4	32.4	16.4	
pm15	3\4	164.4	−7.1	20.8	27.7	
Mean	14\15	178.8	8.5	21.4	8.8	179.2, −3.8
Pole						51.3°N, 71.7°E
Upper member, north Utah (40.57°N, 109.58°W)						
pm17	4\4	191.2	−8.5	59.9	12	
pm18	4\4	176.8	1.2	98.1	9.3	
pm19	4\4	182.2	15.3	68.4	11.2	
pm20	4\4	163.8	9.5	265.8	5.6	
pm21	4\4	170.6	15.2	185.1	6.8	
pm22	4\4	200.9	8.3	73.6	10.8	
Mean	6\6	180.9	7.0	25	13.7	181.0, −5.6
Pole						52.2°N, 68.8°E
Combined	20\22					51.6°N, 70.9°E
Gartra Member, north Utah (40.93°N – 109.9°W)						
cb2	5\5	329.6	64.4	406.8	3.8	
cb3	5\5	328.4	56.3	278.0	4.6	
cb4	6\6	334.3	34.4	840.9	2.3	
cb5	5\5	351.5	61.2	431.8	3.7	
cb6	6\6	320.0	53.7	556.6	2.8	
cb7	5\5	326.3	47.8	467.5	3.5	
Mean	6\7	331.2	53.4	43.3	10.3	341.8,12.9
Pole						52.0°N – 100.3°E
Chinle-Nugget Transition (40.86°N, 109.7°W)						
sc1	7\7	329.9	77.3	38.1	9.9	
sc2	8\8	195.5	−59.4	24.2	11.5	
sc3	8\8	352.6	46.8	143.7	4.6	
sc4	7\7	188.6	−68.7	53.5	8.3	
sc5	5\6	5.9	53.5	13.3	21.8	
Mean	5\5	1.3	61.7	34.0	13.3	192.6, −6.8
Pole						50.9°N, 50.1°E

[a]Here $n\backslash n_d$ is the number of sampled usedνmber of samples demagnetized; D and I are the declination and inclination, respectively; k is Fisher's precision parameter, α_{95} is the radius of the confidence interval around the mean.

removed with AF inductions of ~30 mT in samples from the Moenave and Rock Point formations, thus confirming the presence of maghemite or magnetite. Wingate sandstones and the rocks from strata sampled in the Uinta Mountains show little evidence of a low coercivity phase.

[32] Thermal demagnetization of a three component IRM, using inductions of 3.0, 0.5, and 0.12 T [*Lowrie*, 1990], further clarify the role of ferrimagnetic phases. These data indicate that coarse-grained magnetite and maghemite ($IRM_{0.12T}$) contribute significantly to the IRM and by inference to the natural remanence. The $IRM_{0.5T}$ component is negligible, suggesting the absence of cubic phases of finer grain sizes, and hence higher coercivities. Thus the large percentage of the (IRM_{3T}) removed with AF inductions of 100 to 150 mT likely resides in hematite. In samples of the Moenave Formation a small fraction the $IRM_{0.12T}$ component remains after heating to 600°C (Figure 9d) indicating the presence of maghemite. The relatively lower unblocking temperatures of the hematite in samples of the Rock Point Formation and Comb Ridge is also evident in Figures 9e and 9f, as is the relatively small contribution of a cubic phase in the Wingate Sandstone.

[33] Above, we suggested that a cubic phase might carry a secondary overprint. The three-component IRM demagnetization experiments support a viscous origin of this overprint by providing strong evidence for multidomain magnetite and maghemite. Nonetheless, inductions of 100 mT AF are insufficient to remove all the north directed overprint. Part

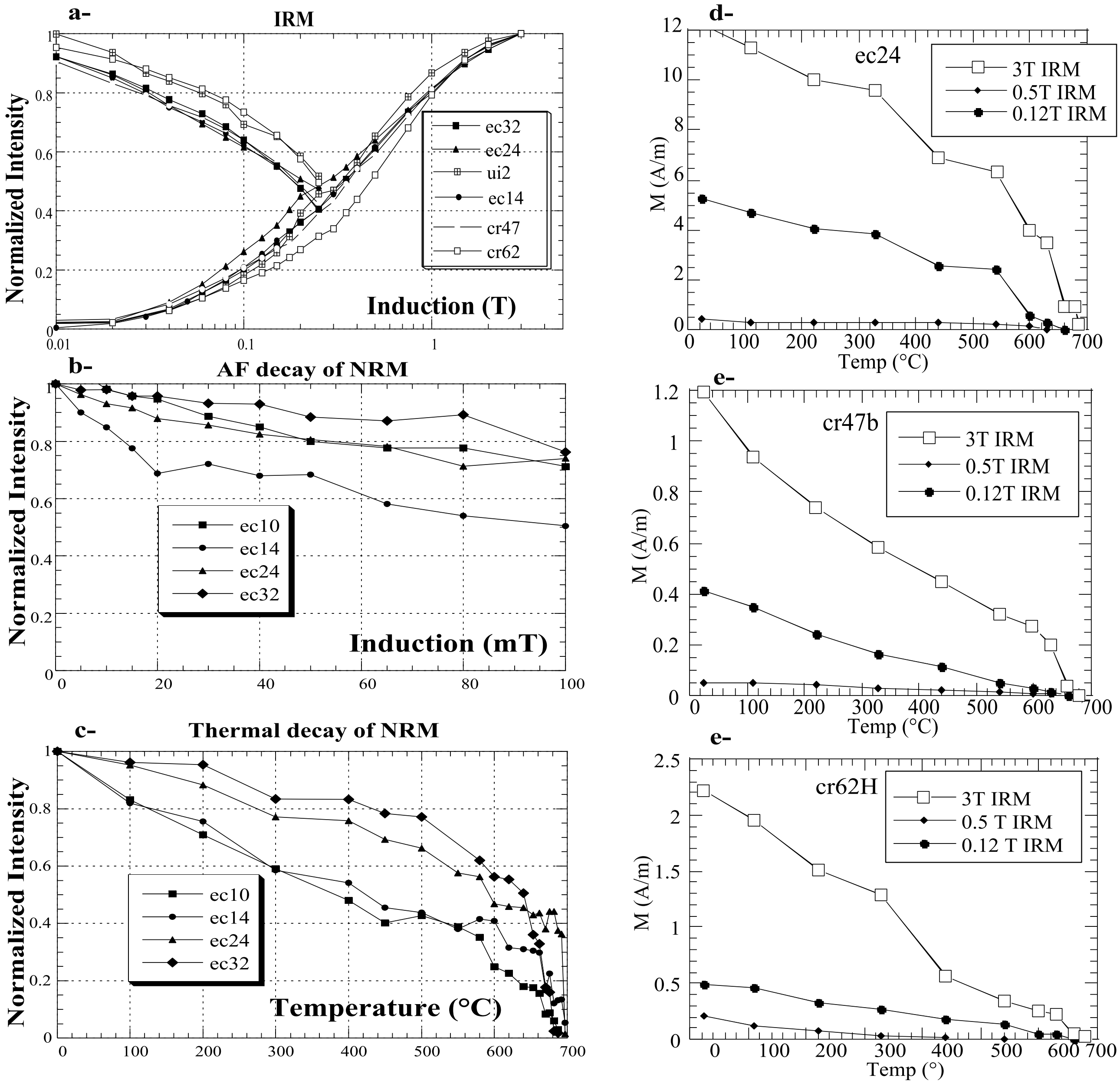

Figure 9. (a) Isothermal remanence acquisition for selected samples, (b) decay of the NRM for samples of the Moenave Formation demagnetized by AF, and (c) samples demagnetized thermally. (d–f) Three-component IRM thermal demagnetization experiments. Sample ec24 is from the anomalous declination interval in the upper Dinosaur Canyon Member. Samples cr47 and cr62 are for the Rock Point Formation and Wingate Sandstone, respectively, at Comb Ridge.

of this overprint must clearly reside in hematite. The steep overprint is, therefore, a composite of a viscous remanence, residing in magnetite and maghemite, and a chemical remanence residing in hematite; both are of relatively recent origin.

[34] Sampled strata of the Moenave and Wingate formations are dominantly fine sandstones to coarse siltstones. Sand/silt grains are subrounded to subangular (Figure 10) and well sorted. Detrital oxide grains are very rare. The few detrital oxide grains identified are small (less than a few tens of microns) and are interpreted to have been specular hematite grains; there is no evidence that the grains were originally magnetite grains deposited and then oxidized in situ to hematite. Most hematite is in the form of poorly developed, spatially very irregular coatings on grains or as alteration products of rare detrital ferromagnesian silicate grains (Figure 10). The relatively low NRM intensities of most of the Moenave and Wingate strata are consistent with a poorly developed authigenic hematite grain coating in these rocks. Because little, if any, of the iron oxides are of detrital origin, we infer that the remanence is not of detrital origin and is not affected by inclination shallowing during deposition.

[35] In contrast, detrital material is present in most of the samples collected in the north flank of the Uinta Mountains. The detrital origin of the remanence in Gartra Member strata is evident in that specular hematite grains form heavy-mineral laminations. The grains are small (mostly less than

10 microns in diameter) and are thoroughly oxidized; primary features are difficult to recognize.

5. Magnetic Stratigraphy

[36] The use of red beds for magnetostratigraphy has been widely debated in the literature. The debate centers on the timing of remanence acquisition [*Larson and Walker*, 1985]. Two lines of evidence suggest that the ChRM of Glen Canyon and Chinle strata provides a reliable record of the ancient geomagnetic field direction, and thus, may be used to construct a magnetic stratigraphy across the Triassic/Jurassic boundary. First, the shallow inclinations are consistent with shallow equatorial paleolatitudes for latest Triassic/earliest Jurassic time, derived from independent studies. Second, there is clear evidence of layer-parallel magnetic polarity zonation. The presence of reverse polarity strata in the upper Dinosaur Canyon Member is consistent with observations in correlative strata in the Vermilion Cliffs [*Ekstrand and Butler*, 1989]. Normal polarities in the thick sandstone bed at site cr51 in a road cut of the Rock Point Formation at Comb Ridge are consistent with normal polarities in a similar bed in a natural outcrop (site cr49) about 5 km north of the highway section. However, data for sites in the Comb Ridge section, as well as petrographic observations, indicate that the lock-in mechanism is not compatible with a depositional remanence (DRM). Rather, the fact that both polarities are observed in the same layer and several samples contain spurious intermediate directions is consistent with slow acquisition of a chemical remanence (CRM). The timing of remanence lock-in cannot be established by field tests, but it seems fast enough to record relatively short polarity events.

[37] At Comb Ridge, the Rock Point Formation and the lower half of the Wingate Sandstone ($\sim$100 m of section) contain magnetizations of nearly uniform reverse polarity (Figure 11). A short interval of normal polarity is recorded in the upper part of the Rock Point Formation, and the upper half of the Wingate section (most of the former Lukachukai Member) is normally magnetized. The polarity stratigraphy at Comb Ridge (Figure 11) is analogous to that observed in the Redonda Formation in eastern New Mexico [*Reeve and Helsley*, 1972], with which the Rock Point Formation has been correlated on the basis of litho- and biostratigraphic data [*Lucas*, 1991, 1993]. A correlation of the Redonda polarity sequence with the Rock Point-Wingate sequence is permissible (thin solid lines, Figure 11).

[38] The magnetic polarity zonation recorded in the southern Uinta Mountains, in strata of the red siltstone and upper members, further supports the presence of a long reverse polarity magnetozone during much of Rock Point time. Magnetic polarity is uniformly reverse in strata of the red siltstone member, and the upper member contains a short normal polarity event within a predominantly reverse polarity magnetozone. The correlation between the section in the Uinta Mountains and the Comb Ridge section is tenuous, because the sampling resolution is low. Nonetheless, we note the presence of a short normal polarity interval that is nearly coincident with the onset of eolian deposition in both sections. This is consistent with the lithostratigraphic correlation of *Lucas* [1993], who equated the upper member and red siltstone member in the Uintas with the Rock Point Formation.

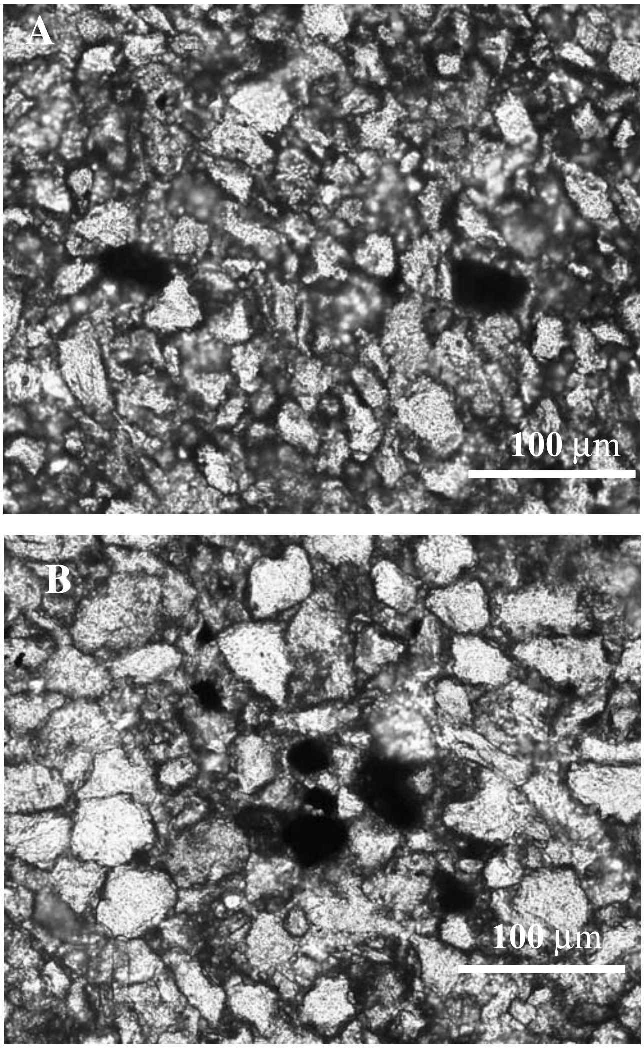

Figure 10. Transmitted light photomicrograph of a typical sample of (a) the Moenave Formation and (b) the Wingate Sandstone.

[39] As noted by *Lucas and Heckert* [2001], fossil evidence indicates a Late Triassic (Rhaetian) age for much if not all of the Wingate Sandstone (Figure 2). As mentioned above, there is no evidence that any part of the Wingate (Lukachukai) is Jurassic. Indeed, it has a Triassic phytosaur at its base and Triassic footprints (esp. *Brachychirotherium*) higher up, essentially to its top (Figure 2). We must thus seek correlation of the Wingate-Rock Point polarity sequence with other latest Triassic polarity stratigraphies.

[40] A standard, globally documented, magnetic polarity stratigraphy for the Late Triassic-Early Jurassic does not yet exist, but high quality data are available for nonmarine strata of the Newark Supergroup [*Kent and Olsen*, 1999; *Kent et al.*, 1995], and limited data are available for marine sequences in Europe and South America. Clearly, correlation of the polarity zonation observed in the upper Chinle Group with the polarity sequence proposed for the Rhaetian interval in the Newark Basin is not unique, nor clear-cut (Figure 11). There are no diagnostic patterns in the upper Chinle Group that can be used for direct correlation, and our sampling scheme, dictated by the availability of suitable

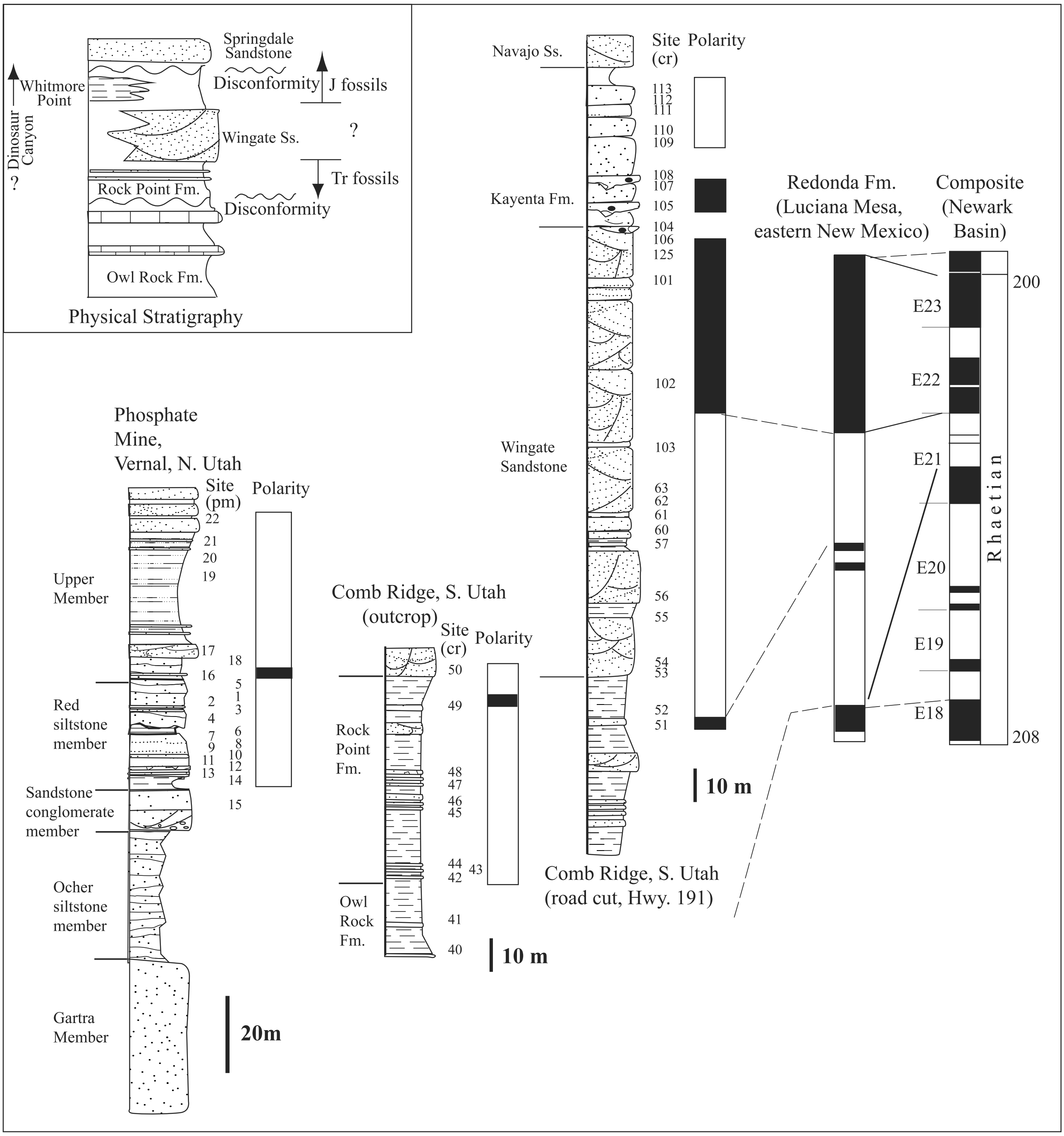

Figure 11. Schematic stratigraphic sections studied, showing sampling levels and associated polarity stratigraphy. For comparison, we show the polarity sequence of the Redonda Formation in eastern New Mexico [*Reeve and Helsely*, 1972] and the Newark composite for the latest Triassic [*Kent et al.*, 1995].

materials, yielded a polarity stratigraphy of comparatively lower resolution than the Newark sequence.

[41] The Rhaetian interval in the Newark basin is dominantly of reverse polarity, although it contains ten short events of normal polarity [*Kent et al.*, 1995]. This general pattern is also supported by data for the upper Blomidon Formation in Nova Scotia [*Kent and Olsen*, 2000], but bears little resemblance to what we observed in the upper Chinle Group. Dashed lines in Figure 11 show possible correlations of the southwest NA magnetostratigraphy to the Newark composite, based on the assumption that the upper Newark and the upper Chinle Group span the same time interval. This correlation would allow the long reverse polarity magnetozone in the upper Chinle Group to correspond to magnetozones E18 to E21 of the Newark composite sequence and the long normal polarity interval in the upper Wingate to magnetozones E22 and E23. Assuming this correlation is valid, one must conclude that either the

southwest NA section contains several hiatuses, or that the sampling interval was insufficient to describe the polarity sequence recorded. It is also possible that the remanence acquisition process is much longer in duration than the average length of a Late Triassic magnetozone, but the above discussion suggests that this is unlikely.

[42] We prefer an alternative correlation. Although the sampling interval is indeed of low resolution, we argue that the reversal frequency observed in the Newark sequence is so high that it is simply unlikely our sampling scheme missed so many reversals. Clearly, additional reversals may exist in the less resistant hematitic mudstone and claystone intervals that we could not sample, but probably not enough to change the overall zonation. Moreover, there is little evidence to support the presence of multiple hiatuses in the sections sampled. The Rock Point-Wingate contact is transitional and the section lacks paleosols or conglomerate-bearing horizons that would indicate long periods of non-deposition or erosion. We propose instead that the entire Rock Point-Wingate section was deposited over a relatively short time period, most likely equivalent to only two of the polarity zones of the Newark sequence. This correlation (thick solid lines in Figure 11) associates the uppermost Rhaetian magnetozones E21 to E23 of the Newark sequence to the sections from the American Southwest.

[43] This interpretive correlation results in several allowable consequences. Overall, it is possible that specific Mesozoic continental sedimentary sequences preserved in the American Southwest were deposited over relatively short periods of time, and thus that only parts of the geomagnetic polarity record over any interval of time are well recorded. The preservation of sediment deposited in continental environments is a result of complex interaction among geomorphic, tectonic and climatic processes. Over 400 km separate the Redonda and Comb Ridge sections, in eastern New Mexico and southeast Utah, respectively. A similar distance separates the Comb Ridge and Vernal sections. The good correlation of polarity sequences across such a large distance of the American Southwest suggests that a regional-scale event is responsible for the preservation of these deposits. Rather than speculate about a specific event that favored preservation of these uppermost Triassic sections, we argue below that the fortuitous preservation of strata of this age facilitates the recording of the J1 cusp of the North American APWP.

[44] The Moenave Formation is almost entirely of normal polarity (Figure 12). A thin zone of reverse polarity exists in the upper Dinosaur Canyon Member, about six meters below the base of the Springdale Sandstone. Field evidence suggests that the Springdale-Dinosaur Canyon contact at the Echo Cliffs is a disconformity. There is no change in polarity at the contact (i.e., between sites ec34 and ec35); thus the presence of an important hiatus at the contact cannot be confirmed.

[45] The inferred intertonguing relation between the Dinosaur Canyon Member of the Moenave Formation and the Wingate Sandstone [*Riggs and Blakey*, 1993] cannot be fully confirmed by the magnetostratigraphy alone. However, recent stratigraphic investigations (S. G. Lucas, unpublished data, 2001) confirm the work of *Harshbarger et al.* [1957], who indicated that the Wingate Sandstone is laterally equivalent to part of the lower Dinosaur Canyon Member of the Moenave Formation. On this basis, the normal polarity zone high in the Wingate almost certainly is equivalent, at least in part, to the normal polarity zone in the lower part of the Moenave section. The short reverse polarity zone high in the Dinosaur Canyon Member is readily correlated to the reverse polarity zone at the base of the Hettangian in the Newark sequence, which is consistent with biostratigraphic evidence indicating that the Triassic/Jurassic boundary is in the lower part of the Moenave Formation. Magnetic polarity is uniformly normal across the Wingate (Lukachukai Member)-Kayenta disconformity at the Comb Ridge section, as it is through the Dinosaur Canyon Member-Springdale Sandstone contact at the Echo Cliffs.

[46] The magnetic polarity stratigraphies for Lower Jurassic marine strata from the Paris basin [*Yang et al.*, 1996] and the Neuquén basin [*Iglesia Llanos and Riccardi*, 2000] provide an opportunity for more specific correlations. As in the Newark basin, Hettangian strata in the Paris basin are predominantly of normal polarity (Figure 12). However, biostratigraphic control in the Paris basin is poor; the Triassic-Jurassic boundary appears to be a hiatus, and considerable uncertainty exists in the position of the Hettangian-Sinemurian boundary [*Yang et al.*, 1996]. The record for the Neuquén basin, in western central Argentina, contains good biostratigraphic control but requires correlation between regional ammonite zonation (Andean) and the European Standard zonation. Hettangian strata of the Neuquén basin are predominantly of reverse polarity. This observation is difficult to reconcile with results for the Newark and Paris basins. *Iglesia Llanos and Riccardi* [2000] suggest that the Hettangian records from the Newark and Paris basins may represent only the lowermost Hettangian. Paleomagnetic data for the Moenave Formation cannot resolve the discrepancies between the Neuquén and Paris sections. The available data suggest that, although there are Jurassic fossils in the upper Dinosaur Canyon Member, of Hettangian age, basal Dinosaur Canyon strata may be of Triassic age. We argue that parts of the thick Moenave section preserved in the Southwest may also represent an interval of relatively short duration, much of which is in the lower Hettangian. Thus, on the basis of limited biostratigraphic data and the magnetostratigraphic correlations presented here, we place the Triassic-Jurassic boundary at the short reversal in the upper Dinosaur Canyon Member (Figure 12).

6. Discussion

6.1. Paleomagnetic Poles

[47] The Moenave Formation at the Echo Cliffs yields a well-defined pole at 63.7°N, 59.0°E (dp = 2.7°, dm = 5.2°, N = 28 sites). This pole lies at an angular distance of $\sim$6° from the Moenave pole determined by *Ekstrand and Butler* [1989], which falls at 58.2°N, 51.9°E; the poles differ at the 95% confidence level (Figure 13a). Twenty-eight and 23 statistically acceptable sites define each of the poles obtained for the Moenave Formation, and in both studies between-site dispersion suggests adequate averaging of paleosecular variation. The pole obtained for sites in the Echo Cliffs is based exclusively on normal polarity directions, whereas the pole obtained in the Vermilion Cliffs includes 4 reverse polarity sites. The higher latitude of the

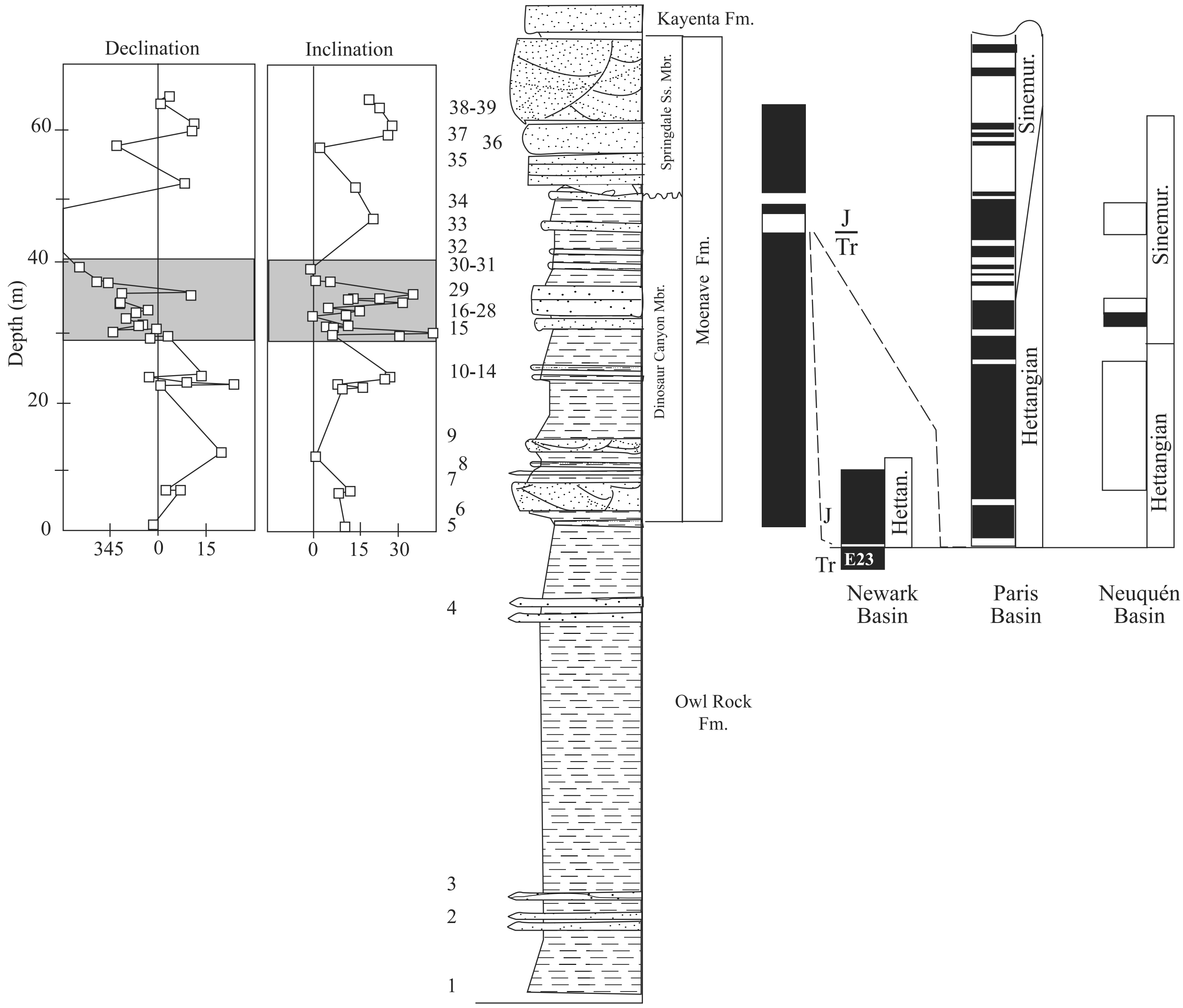

Figure 12. Schematic stratigraphic section of the Moenave Formation at the Echo Cliffs, showing sampling levels and associated polarity stratigraphy. We also show the declination and inclination of individual sampling sites and highlight an interval of ''anomalous'' directions in the Dinosaur Canyon Member. For comparison, we show the magnetostratigraphy of the Newark Extrusive series [*Kent et al.*, 1995], the Paris Basin [*Yang et al.*, 1996], and the Neuquén Basin [*Iglesia Llanos and Riccardi*, 2000].

Echo Cliffs pole may be explained by a small bias in the direction of the recent dipole field. As discussed above, a steep north directed overprint in lower Dinosaur Canyon strata is occasionally stable up to temperatures of 665°C complicating isolation of the ChRM and possibly causing some contamination, but the overall mean of sites with a well developed overprint is not steeper than for sites with a small overprint of lower unblocking temperatures. Other explanation for the higher inclination of the Echo Cliffs is the fact that inclinations in the Springdale Sandstone are higher than in the underlying Dinosaur Canyon and Whitmore Point members; sampling in the Vermillion Cliffs was limited to these two members [*Ekstrand and Butler*, 1989].

[48] At a finer level of detail, we note that the distribution of site-means obtained in our study is somewhat anomalous. A group of 16 site means, sites ec15 to ec32 in the upper strata of the Dinosaur Canyon Member, display (in situ) NNW declinations. The sites are relatively closely spaced in a ~15 meter thick interval (Figure 12) and their mean direction differs by ~15° from the nominal direction observed in sites collected in the lower Dinosaur Canyon Member or the Springdale Sandstone (Table 1). When averaged, the 16 sites with NNW declinations in the upper Dinosaur Canyon Member give a paleomagnetic pole at 64.4°N, 73.6°E (Figure 13a). Poles calculated for the Dinosaur Canyon Member and Springdale Sandstone fall west of this position, it thus appears that the pole backtracks for a distance of ~15°.

[49] The observation of apparent backtracking in the Echo Cliffs sequence of poles is most puzzling, and we have no fully satisfactory explanation for this observation. Systematic APW backtracking was apparently not observed in an earlier study of the Moenave Formation [*Ekstrand and Butler*, 1989], although stratigraphic control was not available for all of the sites reported. The upper Dinosaur Canyon sites do not show particularly unusual demagnet-

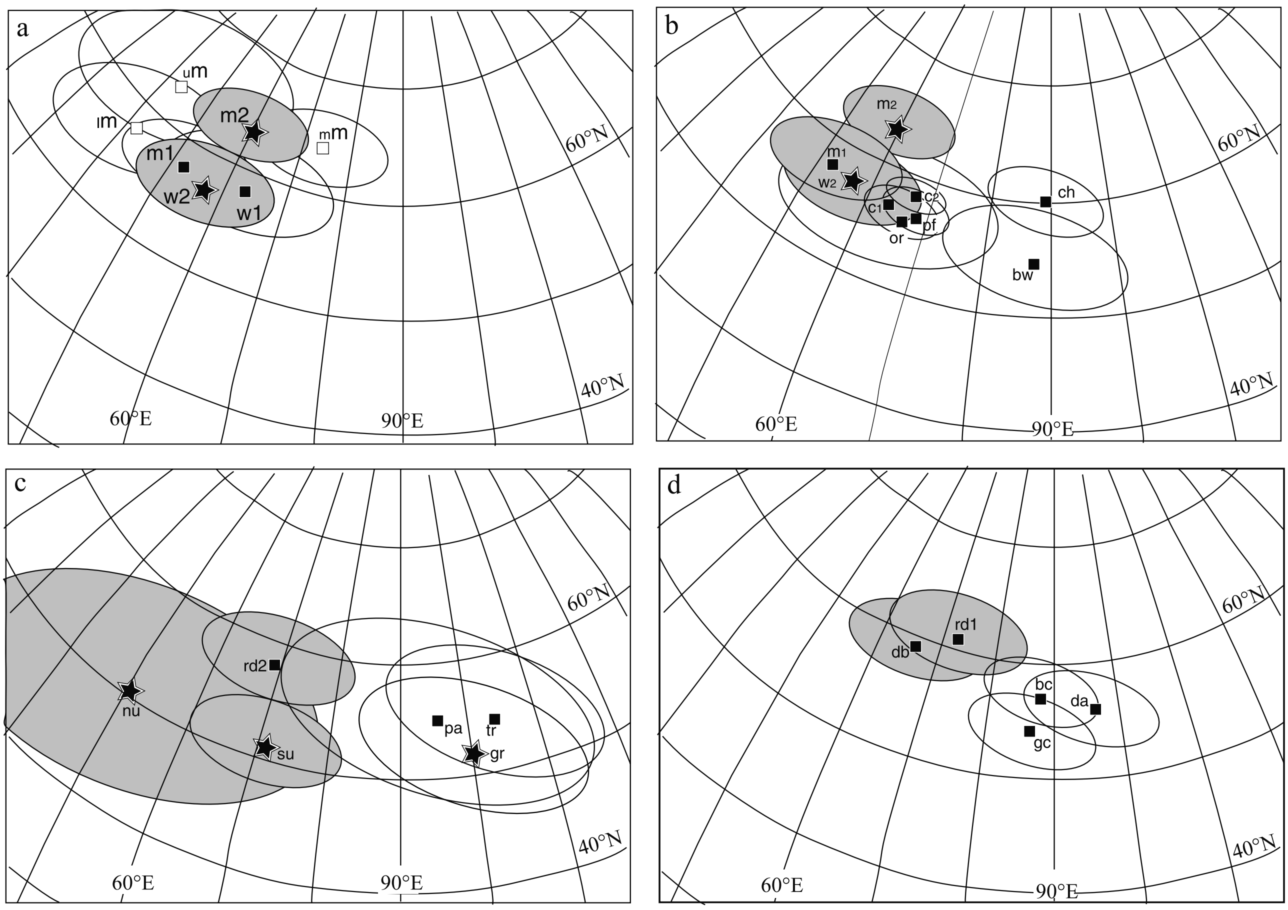

Figure 13. (a) Paleomagnetic poles for the lower Glen Canyon Group. Pole symbols are m1, Moenave Formation [*Ekstrand and Butler*, 1989]; m2, Moenave Formation (this study); w1, Wingate Sandstone [*Reeve*, 1975]; w2, Wingate-Rock Point sequence (this study); lm, pole for sites in the lower Moenave Formation (sites ec6–14); mm, pole for sites in the middle Moenave Formation (ec15–32); um, pole for sites in the upper Moenave Formation (ec34–39). (b) Late Triassic to Hettangian poles for the Colorado Plateau. Pole symbols are bw, Bluewater Creek [*Molina-Garza et al.*, 1998]; ch, lower Chinle Beds, New Mexico [*Molina-Garza et al.*, 1991]; pf, Petrified Forest [*Steiner and Lucas*, 2000]; or, Owl Rock Formation [*Bazard and Butler*, 1991]; c1, Church Rock Formation [*Kent and Witte*, 1993]; c2, Church Rock Formation [*Reeve*, 1975]. (c) Late Triassic poles for the Tertiary uplifts bordering the Colorado Plateau. Symbols are gr, Gartra member, Uinta Mountains (this study); pa, Popo Agie [*Van der Voo and Grubbs*, 1977]; tr, Trujillo and Bull Canyon, Sangre de Cristo Mountains [*Molina-Garza et al.*, 1996]; rd2, Redonda Formation, Sangre de Cristo Mountains [*Molina-Garza et al.*, 1996]; su, Red siltstone and upper members, southern Uinta Mountains (this study); nu, Chinle-Nugget Transition, northern Uinta Mountains (this study); (d) Late Triassic poles for the southern High Plains of the craton interior. Symbols are da, Dockum Group A component [*Molina-Garza et al.*, 1995]; gc, Garita Creek Formation [*Molina-Garza et al.*, 1996]; bc, Bull Canyon Formation [*Bazard and Butler*, 1991]; rd1, Redonda Formation [*Reeve and Helsely*, 1972]; db, Dockum Group B component [*Molina-Garza et al.*, 1995]. Ellipses are 95% confidence intervals.

ization behavior, nor are their rock magnetic properties different from those of other sites (Figures 5 and 9). On the basis of the magnetostratigraphy and the assumption that the Moenave section is partially Rhaetian and partially Hettangian, which stretches out the duration of Moenave deposition, the inferred rate of APW suggested by the pole positions seems too fast to represent actual plate motion.

[50] We interpret this section of the Moenave Formation to have recorded a period of high-amplitude secular variation. Although tenuous, support for this interpretation exists in the similar behavior observed in the Hettangian extrusive series of the Newark Supergroup [*Prévot and McWilliams*, 1989]. They noted that the intermediate volcanic sequence, including the Deerfield basalt, the Holyoke basalt, and the base of the second Watchung basalt, yielded directions that differ by more than 30° from adjacent flows. Directions in the intermediate flows are more easterly and shallow than in the adjacent units. *Prèvot and McWilliams* [1989] suggested that this volcanic sequence records a short-lived, high-amplitude secular variation event. Such interpretation for

VGPs derived from volcanic rocks is not problematic because lava flows are known to contain instantaneous records of the geomagnetic field direction, but the fact that the high-amplitude event is observed over a wide region of northeast North America is inconsistent with a relatively ''instantaneous'' event (a few thousand years in duration). This would imply instantaneous emplacement of these lavas across the entire volcanic province. Prèvot and McWilliams argument for a short-lived event is based on the absence of ''anomalous'' directions in intrusive suites of the Newark Supergroup, but recently, *McEnroe and Brown* [2000] report distinct directions for dike subsets of Early Jurassic age in New England and *Kodama et al.* [1994] report a pole similar to that of the intermediate volcanic flows for sediments baked by a diabase intrusion in the Culpeper basin. The high dispersion of earliest Jurassic poles is well documented in North America, and is also evident in Europe [*Sichler and Perrin*, 1993]. As in the case of the Newark extrusive series, dike subsets reported by *McEnroe and Brown* [2000] are grouped according to their geochemistry. By inference, they are interpreted as temporally distinct groups. High-amplitude secular variation of the geomagnetic field and rapid magma emplacement may combine to produce relatively low dispersion within flow sequences (or within dike subsets), but high dispersion between flows (dikes). Our knowledge about secular variation in the Mesozoic is so rudimentary, that it is also possible that not only the amplitude but also the periodicities of secular variation might have been larger during the episode recorded by the Newark lavas.

[51] There are no accurate means to estimate the duration of the ''short-lived field event'' in the Newark basin; let alone fully establish whether the event is the same event recorded in the Moenave Formation. The interval recording the apparent backtracking episode in the Echo Cliffs section is several meters in thickness and contains several sandstone beds intercalated with mudstone intervals. The interval is considerably thicker than expected for a ''short-lived'' geomagnetic event; yet, the event occurs in a single magnetozone. Although we have interpreted the backtracking episode in Moenave strata to reflect an event similar to the one recorded in the Newark Extrusive series, we have no geologic or statistical reason to exclude these sites from the overall mean.

[52] Alternative explanations for the anomalous directions or backtracking episode in the Dinosaur Canyon Member call for deviation of the time averaged field from the geocentric axial dipole and/or true polar wander, but these are difficult to quantify. As evidence supporting geologically instantaneous emplacement of earliest Jurassic flood basalts in the central Atlantic continues to mount [*Hames et al.*, 2000], we speculate that such an extraordinary geodynamic event could cause true polar wander, and/or geodynamo instability. Although plume magmatism has not been directly linked to true polar wander, upper mantle overheating has [*Gordon*, 1987]. The assumed synchrony of what we infer to be relatively rapid backtracking of the pole during deposition of the Moenave Formation and the formation of the Central Atlantic Magmatic Province provides a provocative link, but one that needs to be explored further.

[53] We combined the Comb Ridge data into a single pole that falls at 57.4°N, 56.6°E (dp = 3.3°, dm = 6.5°; N = 17 sites). All sites with acceptable statistics were combined because the Rock Point Formation and Wingate Sandstone were sampled in a section that does not contain any significant stratigraphic breaks, and because individual poles cannot be precisely determined with six and eleven accepted sites, respectively. The combined pole is near the Wingate pole reported by *Reeve* [1975], which falls at 59°N, 63°E. Our pole supercedes the pole of *Reeve* [1975], which is based on very limited data with complex demagnetization behavior.

[54] Three reliable poles exist for basal Glen Canyon Group strata, two for the Moenave Formation and one for the Rock Point Wingate sequence. The poles span the Rhaetian-Hettangian interval and no clear temporal sequence of poles can be established by superposition or other means. The Moenave pole obtained at the Vermillion Cliffs [*Ekstrand and Butler*, 1989] is based principally on results from strata from the Whitmore Point Member, and therefore is arguably younger than the Wingate pole, based on lithostratigraphic observations. The Moenave pole from Vermillion Cliffs lies west of the Wingate pole, but the Moenave pole obtained at the Echo Cliffs lies east and north of it.

[55] Data from the north flank of the Uinta Mountains define an early Late Triassic (late Carnian) pole, based on samples collected in the Gartra Member, falling at 52.0°N, 100.3°E (dp = 5.4°, dm = 10.5°; N = 6 sites). The pole is derived from a small number of sites; the data are of good quality, but their use in tectonic interpretations is cautioned. The same must be said for a latest Late Triassic (Rhaetian) pole defined by five sites collected across the Chinle-Nugget Sandstone transition. The Rhaetian pole is at 50.9°N, 50.1°E (dp = 8.8°, dm = 17.5°). Data collected in stratigraphic succession on the south flank of the range define two Rhaetian poles, but as in the case of the Comb Ridge section, the poles are statistically indistinguishable. Combined, Rhaetian strata of the southern flank of the Uinta Mountain section give a pole at 51.6°N, 70.9°E (N = 20, dp = 3.5°, dm = 6.9°). This pole, like those obtained for the Comb Ridge and the Echo Cliffs sections, is interpreted to record accurate relative motion of the region with respect to the rotation axis.

6.2. CP Rotation and Rate of APW

[56] Two issues complicate our current understanding of the Late Triassic/Early Jurassic segment of the North American apparent polar wander path and the cusp defined near the end of the Triassic. These are the magnitude of Colorado Plateau rotation and the discrepancy in position between presumably coeval poles from areas in northeast North America and the American Southwest. These fuel a controversy concerning Triassic apparent polar wander and the rate of pole motion [e.g., *Kent and Witte*, 1993]. The locations of the Moenave and Wingate poles are consistent with previous interpretations of large magnitude, fast APW during the Late Triassic and Early Jurassic, as they differ from poles for underlying strata of the Painted Desert Member of the Petrified Forest Formation, and the Owl Rock and Church Rock formations (Figure 13b).

[57] Late Triassic and Early Jurassic data for the American Southwest (Table 4) include over 20 poles from more than a dozen independent studies, all of which used current

Table 4. Summary of Late Triassic-Early Jurassic Paleomagnetic Data From the American Southwest: Chinle Group and Glen Canyon Group[a]

Pole	Location	Statistics				
	Lat °N, Long °E	*N*	*n*	α_{95}	Paleolatitude, °N	Reference
			Colorado Plateau			
Early Jurassic						
Kayenta 2, Arizona	59.0, 66.6	23		2.4°	6.1	1
Kayenta, Utah	61.9, 74.4	7		6.8°	8.9	2
Moenave 2, Arizona	63.7, 59.7	28		5.8°	11.1	3
Moenave 1, Utah	58.2, 51.9	23		4.5°	6.5	4
Late Norian-Rhaetian						
Wingate-Rock Point, Utah	57.4, 56.6	17		6.4°	5.2	3
Church Rock 1, Utah	57.5, 63.3	9		10.7°	4.7	5
Church Rock 2, Utah	59.0, 67.0		28	2.5°	1.2	14
Owl Rock, Arizona	56.5, 66.4	18		3.4°	3.6	6
Early Norian						
Painted Desert, Petrified Forest, Arizona	58.4, 68.5		24	4.3°	5.4	7
Late Carnian						
BlueMesa, Petrified Forest, Arizona	56.5, 68.2		41	4.0°	3.3	7
Bluewater Creek, New Mexico	55.2, 87.5	13		6.7°	3.5	8
Chinle lower beds, New Mexico	60.5, 88.9	17		5.0°	8.6	10
			Plateau: Deformed Bordering Region			
Early Jurassic						
Late Norian-Rhaetian						
Upper Chinle, Northern Uintas, Utah	50.9, 50.1	6		17.5°	3.4	3
Redonda Formation, New Mexico	58.6, 68.9	8		6.9°	5.6	9
Upper Chinle, Southern Uintas, Utah	51.6, 70.9	20		6.9°	−1.4	3
Early Norian						
Trujillo-Bull Canyon, New Mexico	54.9, 104.1	9		9.7°	6.2	9
Late Carnian						
San Rosa-Garita Creek, New Mexico	53.1, 88.8	20		6.9°	1.4	9
Lower Chinle, Northern Uintas, Utah	52.3, 100.2	6		10.3°	2.9	3
Popo Agie, Wyoming	55.5, 95.5		25	14.0°	4.8	13
			Plateau: Undeformed Surrounding Region			
Early Jurassic						
Dockum-B complex, Texas	59.7, 65.7	14		6.0°	6.8	12
Late Norian-Rhaetian						
Redonda Formation, New Mexico	61.4, 72.4	2		6.0°	8.4	11, 1
Early Norian						
Bull Canyon, New Mexico	57.4, 87.8	15		5.0°	5.5	1
Late Carnian						
Garita Creek, New Mexico	54.0, 86.7	12		5.7°	2.0	9
Dockum A, Texas	56.4, 96.3	12		6.6°	5.8	12

[a]Here *N* (*n*) is the number of sites (samples) used in pole calculations. The confidence region (α_{95}) of the formation mean is listed for reference. Paleolatitude is for 37°N–100°W. References: (1) *Bazard and Butler* [1991]; (2) *Helsely and Steiner* [1974]; (3) this study; (4) *Ekstrand and Butler* [1989]; (5) *Witte and Kent* [1993]; (6) *Bazard and Butler* [1991]; (7) *Steiner and Lucas* [2000]; (8) *Molina-Garza et al.* [1998]; (9) *Molina-Garza et al.* [1996]; (10) *Molina-Garza et al.* [1991]; (11) *Reeve and Helsley* [1972]; (12) *Molina-Garza et al.* [1995]; (13) *Van der Voo and Grubbs* [1977]; (14) *Reeve* [1975].

laboratory and data analysis techniques and pass modern reliability criteria. Data are arranged in four age groups; no numerical ages exist for rocks studied for this data set, but the temporal sequence is indisputable, as this is afforded by stratigraphic superposition. The only uncertain relative age is for the pole derived from the Dockum Group B overprint, which is based on a secondary magnetization [*Molina-Garza et al.*, 1995]. Poles are listed for three regions of the American Southwest: the Colorado Plateau, the tectonically deformed plateau-bordering region of Cenozoic uplifts (which includes basement uplifts of central New Mexico, Colorado, and northern Utah), and the undeformed cratonic region (which includes flat-lying strata in the High Plains of west Texas and eastern New Mexico).

[58] Three poles from basal Glen Canyon strata are ~5° west of well-defined poles in the Church Rock and Owl Rock formations of the Chinle Group [*Witte and Kent*, 1993; *Bazard and Butler*, 1991], of Norian age, and ~14° west of a more scattered group of late Carnian-early Norian poles (Figure 13b). All of the poles are for localities within the CP, thus rotation is not an issue if the plateau is assumed to have rotated as a rigid body. Together, the intermediate position of the Church Rock, Painted Desert, and Owl Rock poles between both groups, and their stratigraphic position between them, suggest that APW was relatively continuous during the Late Triassic. This data set does not support recent claims of a Late Triassic stillstand of the paleomagnetic pole [e.g., *Steiner and Lucas*, 2000]. A similar conclusion is reached with two independent data sets (Figures 13c and 13d). Poles derived from latest Triassic Chinle Group strata in the deformed margin region of the CP, the Sangre de Cristo and Uinta Mountains, are far removed from a group of late Carnian-early Norian poles from the same regions. Admittedly, this data set is of lesser quality than the CP data set. Nonetheless, a similar separation between late Carnian-early Norian and latest Triassic poles

is evident in the data for essentially undeformed strata on the High Plains of the craton interior, where the poles have smaller associated uncertainties (Figure 13d). A simple APW trend, documenting ~12° of pole motion during Late Triassic time is evident in the High Plains data set; a similar amount of APW is evident from data from strata exposed in the uplifts that border the CP and in the CP itself, although the trends in the CP and bordering region are not as simple as in the cratonic subset.

[59] *Kodama et al.* [1994] reported easterly directions for sediments baked by a diabase intrusion in the Culpeper basin. These data have been interpreted to support the existence of the J1 cusp in the North American APWP; the data also supports the notion of significant apparent polar wander during Late Triassic time. Other record supporting both the cusp and a fast rate of Late Triassic apparent polar wander is available from Brittany, for counterparts of the eastern North American magmatic province [*Sichler and Perrin*, 1993].

[60] The Late Triassic data set for southwest North America also shows a small but consistent northward drift of the region. Paleolatitudes calculated for a central point in the CP (Table 4) indicate that the Monument Valley region of the American Southwest drifted north ~4° from late Carnian (225Ma) to Sinemurian time (~190 Ma), from ~4°N to ~8°N of the paleoequator. Late Triassic paleolatitudes ranging from 1.5°S to 8°N are slightly lower than has been assumed in general models of monsoonal circulation during Chinle deposition (5° to 15°N according to *Dubiel et al.* [1991]).

[61] Although the poles listed in Table 4 span an interval of close to 35 million years, we propose that most of the poles possibly represent "punctual" averages of the pole position over short intervals of time, spanning perhaps 1 or 2 Myr. This assertion stems from the magnetostratigraphic record. Examples for the Bluewater Creek Formation, Trujillo Sandstone, red siltstone member, Rock Point Formation, Wingate Sandstone, Garita Creek Formation, and Dockum Group suggest that intervals several tens of meters thick (often encompassing a complete lithostratigraphic unit) are characterized by a near uniform or uniform polarity zonation. If current estimates of reversal frequency during the Late Triassic are reasonably accurate [e.g., *Gallet et al.*, 1992], near uniform polarity records suggest that the stratigraphic interval sampled represents a short period of deposition preserved due to favorable conditions. This in turn facilitated detailed recording of APW features such as the J1 cusp.

[62] We hypothesize that the discrepancy between Hettangian (lowermost Jurassic) data from the Newark Basin and coeval strata in western North America is possibly because the rocks in northeast NA sampled relatively short intervals during an episode of "anomalous" behavior of the geomagnetic field characterized by high-amplitude secular variation, the same anomalous interval recorded in Moenave strata.

[63] Although no direct measurement is available for the duration of the J1 cusp, the relative consistency of Moenave and Wingate poles, and their relatively short angular distance from latest Triassic (e.g., Church Rock), and early Jurassic (Sinemurian) (e.g., Kayenta) poles, suggest that the cusp is not a geologically instantaneous feature. The cusp possibly spans part of the Rhaetian and part of the Hettangian, or several million years. There is no consensus on the duration of the Rhaetian, but the Triassic/Jurassic boundary is now dated at ~200 Ma [*Palfy et al.*, 2000].

[64] A complete assessment of the CP rotation was presented by *Molina-Garza et al.* [1998]. The data obtained in this study provide further support for their conclusions, but we have not attempted to revise their rotation estimates. In fact, that study demonstrated that pole-to-pole comparisons yield biased estimates and should be viewed with concern. Nonetheless, we can comment on the discussion on CP rotation, based in part on the additional data presented here.

[65] The late Carnian-Early Norian subsets for the craton High Plains and the CP bordering region yield a cluster of poles, all of which are indistinguishable at the 95% level. They were combined to provide an estimate of the reference cratonic pole position for this time (early Late Triassic). The combined mean falls at 55.0°N, 94.2°E (N = 7 poles, K = 363.8, A95 = 3.2°). Similarly, latest Triassic-earliest Jurassic poles for the region give a combined late Late Triassic pole that falls at 57.9°N, 69.5°E (N = 4, K = 315.8, A_{95} = 5.2°; it excludes the pole from the northern Uintas). For similar age poles in the CP, for correlative strata, we obtain a late Carnian-early Norian pole that falls at 58.0°N, 78.2°E (N = 4, K = 154.3, A_{95} = 7.4°) and a latest Triassic-earliest Jurassic pole at 58.8°N, 60.9°E (N = 6, K = 405.1, A_{95} = 3.3°). The relatively good agreement of coeval poles near the time of the J-1 cups suggests that internal deformation of the CP has not contributed to the apparent observation of fast Late Triassic APW and features such as the J1 cusp. At face value, the distance between early Late Triassic poles is 9.3° and the distance between late Late Triassic poles is 4.0°; both are consistent with a small (<5°) clockwise rotation of the plateau, as has been concluded by us and by other workers.

[66] **Acknowledgments.** We thank Aaron Cavosie, Laura Pletsch-Rivera, Mike Petronis, Alfred Gomez, and Viorel Atudorei for their help in the field. The original manuscript benefited from careful reviews by K. Kodama, G. Muttoni, and T. Rolph.

References

Bazard, D. R., and R. F. Butler, Paleomagnetism of the Chinle and Kayenta Formations, New Mexico and Arizona, *J. Geophys. Res.*, *96*, 9847–9871, 1991.

Besse, J., and V. Courtillot, Paleogeographic maps of the continents bordering the Indian Ocean since the Early Jurassic, *J. Geophys. Res.*, *93*, 11,791–11,808, 1988.

Bird, P., Kinematic history of the Laramide Orogeny in latitudes 35°–49°N, western United States, *Tectonics*, *17*, 780–801, 1998.

Bump, A. P., et al., Progressive development of brittle and semi-brittle structures on the Monument Uplift, Utah, and their implications for regional strain, *Geol. Soc. Am. Abstr. Programs*, *32*(5), 4, 2000.

Cather, S. M., Implications of Jurassic, Cretaceous, and Proterozoic piercing lines for Laramide oblique-slip faulting in New Mexico and rotation of the Colorado Plateau, *Geol. Soc. Am. Bull.*, *111*, 849–868, 1999.

Dubiel, R. F., J. T. Parrish, J. M. Parrish, and S. C. Good, The Pangean megamonsoon- evidence from the Upper Triassic Chinle Formation, Colorado Plateau, *Palaios*, *6*, 347–370, 1991.

Ekstrand, E. J., and R. F. Butler, Paleomagnetism of the Moenave Formation: Implications for the Mesozoic North American apparent polar wander path, *Geology*, *17*, 245–248, 1989.

Fisher, R. A., Dispersion on a sphere, *Proc. R. Soc. London, Ser. A*, *217*, 295–305, 1953.

Gallet, Y., J. Besse, L. Krystyn, J. P. Marcoux, and H. Theveniaut, Magnetostratigraphy of the Late Triassic Bolucktasi Tepe section (southwestern Turkey); implications for changes in magnetic reversal frequency, *Phys. Earth Planet. Inter.*, *73*, 85–108, 1992.

Gordon, R. G., Polar wandering and paleomagnetism, *Annu. Rev. Planet. Sci.*, *15*, 567–593, 1987.

Gordon, R. G., A. Cox, and S. O'Hare, Paleomagnetic Euler poles and the apparent polar wander and absolute motion of North America since the Carboniferous, *Tectonics*, *3*, 499–537, 1984.

Gregson, J. D., and E. A. Erslev, Heterogeneous Laramide deformation in the Uinta Mountains, Colorado and Utah, in *Fractured Reservoirs: Characterization and Modeling*, edited by T. E. Hoak, A. L. Klawitter, and P. K. Blomquist, pp. 137–154, Rocky Mt. Assoc. of Geol., Denver, Colo., 1997.

Hames, W. E., P. R. Renne, and C. Ruppel, New evidence for geologically instantaneous emplacement of earliest Jurassic central Atlantic magmatic province basalts on the North American margin, *Geology*, *28*, 859–862, 2000.

Harrison, C. G. A., and T. B. Lindh, A polar wandering curve for North America during the Mesozoic and Cenozoic, *J. Geophys. Res.*, *87*, 1903–1920, 1982.

Harshbarger, J. W., C. A. Repenning, and J. H. Irwin, Stratigraphy of the uppermost Triassic and the Jurassic rocks of the Navajo Country, *U.S. Geol. Surv. Prof. Pap., 291*, 12–26, 1957.

Helsley, C. E., and M. B. Steiner, Magnetic polarity sequence of the Upper Triassic Kayenta Formation, *Geology*, *2*, 191–194, 1974.

Iglesia Llanos, M. P., and A. C. Riccardi, The Neuquén composite section: Magnetostratigraphy and biostratigraphy of the marine lower Jurassic from the Neuquén basin (Argentina), *Earth Planet. Sci. Lett.*, *181*, 443–457, 2000.

Irving, E., and G. A. Irving, Apparent polar wander paths Carboniferous through Cenozoic and the assembly of Gondwana, *Geophys. Surv.*, *5*, 141–188, 1982.

Kent, D. V., and P. E. Olsen, Astronomically tuned geomagnetic polarity timescale for the Late Triassic, *J. Geophys. Res.*, *104*, 12,831–12,842, 1999.

Kent, D. V., and P. E. Olsen, Magnetic polarity stratigraphy and paleolatitude of the Triassic-Jurassic Blomidon Formation in the Fundy Basin (Canada): Implications for early Mesozoic tropical climate gradients, *Earth Planet. Sci. Lett.*, *179*, 311–324, 2000.

Kent, D. V., and W. K. Witte, Slow apparent polar wander for North America in the Late Triassic and large Colorado Plateau rotation, *Tectonics*, *12*, 291–300, 1993.

Kent, D. V., P. E. Olsen, and W. K. Witte, Late Triassic-earliest Jurassic geomagnetic polarity sequence and paleolatitudes from drill cores in the Newark rift basin, eastern North America, *J. Geophys. Res.*, *100*, 14,965–14,998, 1995.

Kirschvink, J. L., The least-squares line and plane and the analysis of paleomagnetic data: Examples from Siberia and Morocco, *Geophys. J. R. Astron. Soc.*, *62*, 699–718, 1980.

Kodama, K. P., M. T. Cioppa, E. Sherwood, and A. C. Warnock, Paleomagnetism of baked sedimentary rocks in the Newark and Culpeper basins: Evidence of the J1 cusp and significant Late Triassic apparent polar wander from the Mesozoic basins of North America, *Tectonics*, *13*, 917–928, 1994.

Larson, E. E., and T. R. Walker, Comment on ''Paleomagnetism of a polarity transition in the lower(?) Triassic Chugwater Formation, Wyoming'' by Emilio Herrero-Bervera and Charles E. Helsley, *J. Geophys. Res.*, *90*, 2060–2062, 1985.

Litwin, R. J., The palynostratigraphy and age of the Chinle and Moenave formations, southwestern U.S.A., Ph.D. dissertation, 282 pp., Pa. State Univ., University Park, 1986.

Lockley, M. G., K. Conrad, M. Paquette, and J. O. Farlow, Distribution and significance of Mesozoic vertebrate trace fossils in Dinosaur National Monument, *Univ. Wyo. Natl. Park Serv. Res. Rep.*, *16*, 64–85, 1992.

Lowrie, W., Identification of ferromagnetic minerals in a rock by coercivity and unblocking temperature properties, *Geophys. Res. Lett.*, *17*, 159–162, 1990.

Lucas, S. G., Correlation of Triassic strata of the Colorado Plateau and southern High Plains, New Mexico, *N. M. Bur. Mines. Miner. Resour. Bull.*, *137*, 47–56, 1991.

Lucas, S. G., The Chinle Group: Revised stratigraphy and chronology of Upper Triassic nonmarine strata in the western United States, *Mus. North. Ariz. Bull.*, *59*, 27–50, 1993.

Lucas, S. G., The thyrophoran dinosaur Scelidosaurus from the Lower Jurassic Lufeng Formation, Yunnan, China, *Mus. North. Ariz. Bull.*, *60*, 81–85, 1996.

Lucas, S. G., and A. B. Heckert, Theropod dinosaurs and the Early Jurassic age of the Moenave Formation, Arizona-Utah, USA., *Neues Jahrb. Geol. Palaontol. Monatsh.*, *2001*(7), 435–448, 2001.

Lucas, S. G., A. B. Heckert, J. W. Estep, and O. J. Anderson, Stratigraphy of the Upper Triassic Chinle Group, Four Corners region, *Field Conf. Guideb. N. M. Geol. Soc.*, *48*, 81–108, 1997.

Lupe, R., and N. J. Silberling, Genetic relationship between Lower Mesozoic Continental strata of the Colorado Plateau and Marine strata of the western Great Basin: Significance for accretionary history of Cordilleran lithotectonic terranes, in *Tectonostratigraphic Terranes of the Circum-Pacific Region*, edited by D. G. Howell, pp. 263–270, Circum-Pac. Counc. for Energy and Miner. Resour., Houston, Tex., 1985.

Marzolf, J. E., Palinspastic reconstruction of early Mesozoic sedimentary basins near the latitude of Las Vegas: Iimplications for the early Mesozoic Cordilleran cratonal margin, in *Mesozoic Paleogeography of the Western United State*, vol. II, *Field Trip Guideb. 71*, pp. 433–462, Pac. Sect., Soc. of Econ. Paleontol. and Mineral., Los Angeles, Calif., 1993.

May, S. R., and R. F. Butler, North American Jurassic apparent polar wander: Implications for plate motion, paleogeography and Cordilleran tectonics, *J. Geophys. Res.*, *91*, 11,519–11,544, 1986.

McEnroe, S. A., and L. L. Brown, Paleomagnetism, rock magnetism and geochemistry of Jurassic dykes and correlative redbeds, Massachusetts, USA, *Geophys. J. Int.*, *143*, 22–38, 2000.

McFadden, P. L., and M. McElhinny, Classification of the reversal test in palaeomagnetism, *Geophys. J. Int.*, *103*, 725–729, 1990.

Molina-Garza, R. S., J. W. Geissman, R. Van der Voo, S. G. Lucas, and S. N. Hayden, Paleomagnetism of the Moenkopi and Chinle Formations in central New Mexico: Implications for the North American apparent polar wander path and Triassic magnetostratigraphy, *J. Geophys. Res.*, *96*, 14,239–14,262, 1991.

Molina-Garza, R. S., J. W. Geissman, and R. Van der Voo, Paleomagnetism of the Dockum Group (Upper Triassic), northwest Texas, Further evidence for the J1 cusp in the North American apparent polar wander path and implications for rate of Triassic apparent polar wander and Colorado Plateau rotation, *Tectonics*, *14*, 979–993, 1995.

Molina-Garza, R. S., J. W. Geissman, S. G. Lucas, and R. Van der Voo, Paleomagnetism and magnetostratigraphy of Triassic strata in the Sangre de Cristo Mountains and Tucumcari Basin, New Mexico, USA, *Geophys. J. Int.*, *124*, 935–953, 1996.

Molina-Garza, R. S., G. D. Acton, and J. W. Geissman, Carboniferous through Jurassic paleomagnetic data and their bearing on rotation of the Colorado Plateau, *J. Geophys. Res.*, *103*, 24,179–24,188, 1998.

Olsen, P. E., C. Koeberl, H. Huber, A. Montanaria, S. J. Fowell, M. Et-Touhami, and D. V. Kent, Continental Triassic-Jurassic boundary in central Pangea: Recent progress and discussion of an Ir anomaly, in Catastrophic Events and Mass Extinctions: Impacts and Beyond: edited by C. Koeberl and K. G. MacLeod, *Spec. Pap. Geol. Soc. Am.*, *356*, 505–522, 2002.

Palfy, J., J. K. Mortensen, E. S. Carter, P. L. Smith, R. M. Friedman, and H. W. Tipper, Timing the end-Triassic mass extinction: First on land, then in the sea?, *Geology*, *28*, 39–42, 2000.

Peterson, F., and G. N. Pipiringos, Stratigraphic relations of the Navajo Sandstone to Middle Jurassic formations, southern Utah and northern Arizona, *U.S. Geol. Surv. Prof. Pap., 1035-B*, 1–34, 1979.

Pipiringos, G. N., and R. B. O'Sullivan, Principal unconformities in Triassic and Jurassic rocks, western interior United States—A preliminary survey, *U.S. Geol. Surv. Prof. Pap., 1035-A*, 29 pp., 1978.

Poole, F. G., and J. H. Stewart, Chinle Formation and Glen Canyon Sandstone in northeastern Utah and northwestern Colorado, *U.S. Geol. Surv. Prof. Pap., 501-D*, 30–39, 1964.

Prévot, M., and M. McWilliams, Paleomagnetic correlation of Newark Supergroup volcanics, *Geology*, *17*, 1007–1010, 1989.

Reeve, S. C., Paleomagnetic studies of sedimentary rocks of Cambrian and Triassic age, Ph.D. thesis, 426 pp., Univ. of Tex., Dallas, 1975.

Reeve, S. C., and C. E. Helsley, Magnetic reversal sequence in the upper portion of the Chinle Formation, Montoya, New Mexico, *Geol. Soc. Am. Bull.*, *83*, 3795–3812, 1972.

Riggs, N. R., and R. C. Blakey, Early and Middle Jurassic Paleogeography and Volcanology of Arizona and adjacent areas, in *Mesozoic Paleogeography of the Western United States*, vol. II, *Book 71,* edited by G. Dunn and K. McDougall, pp. 347–375, Pac. Sect., Soc. of Econ. Paleontol. and Mineral., Los Angeles, Calif., 1993.

Sichler, B., and M. Perrin, New early Jurassic paleopole from France and Jurassic apparent polar wander, *Earth Planet. Sci. Lett.*, *115*, 13–27, 1993.

Steiner, M. B., Paleomagnetism of the Late Pennsylvanian and Permian; a test of the rotation of the Colorado Plateau, *J. Geophys. Res.*, *93*, 2201–2215, 1988.

Steiner, M. B., and S. G. Lucas, A Middle Triassic paleomagnetic pole for North America, *Geol. Soc. Am. Bull.*, *104*, 993–998, 1993.

Steiner, M. B., and S. G. Lucas, Paleomagnetism of the Late Triassic Petrified Forest Formation, Chinle Group, western United States: Further evidence of ''large'' rotation of the Colorado Plateau, *J. Geophys. Res.*, *105*, 25,791–25,808, 2000.

Van der Voo, R., Phanerozoic paleomagnetic poles from Europe and North America and comparisons with continental reconstructions, *Rev. Geophys.*, *28*, 167–206, 1990.

Van der Voo, R., *Paleomagnetism of the Atlantic, Thethys, and Iapetus Oceans*, 411 pp., Cambridge Univ. Press, New York, 1993.
Van der Voo, R., and K. L. Grubbs, Paleomagnetism of the Chugwater redbeds revisited (Wyoming, USA), *Tectonophysics*, *164*, 13–24, 1977.
Wawrzyniec, T. F., J. W. Geissman, M. D. Melker, and M. Hubbard, Dextral shear along the eastern margin of the Colorado Plateau-A kinematic link between the Laramide orogeny and Rio Grande rifting (ca. 80 Ma to 13 Ma), *J. Geol.*, *110*, 305–324, 2002.
Yang, Z., M. G. Moreau, H. Bucher, J. L. Dommergues, and A. Trouiller, Hettangian and Sinemurian magnetostratigraphy from Paris basin, *J. Geophys. Res.*, *101*, 8025–8042, 1996.
Zijderveld, J. D. A., A.C. demagnetization of rocks: Analysis of results, in *Methods in Rock Magnetism and Paleomagnetism*, edited by D. W. Collinson, K. M. Creer, and S. K. Runcorn, pp. 254–286, Elsevier Sci., New York, 1967.

R. S. Molina-Garza, Centro de Geociencias, UNAM Campus Juriquilla, Querétaro 76230, México. (rmolina@geociencias.unam.mx)

J. W. Geissman, Department of Earth and Planetary Sciences, University of New Mexico, Albuquerque, NM 87131, USA.

S. G. Lucas, New Mexico Museum of Natural History and Science, Albuquerque, NM 87104, USA.