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Early Jurassic magnetostratigraphy and paleolatitudes from the Hartford continental rift basin (eastern North America): Testing for polarity bias and abrupt polar wander in association with the central Atlantic magmatic province

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[1] To determine whether the ~200 Ma central Atlantic magmatic province (CAMP) coincides with a normal polarity bias and a purported abrupt change in polar wander at the J1 cusp, we collected samples for paleomagnetic study from 80 sites distributed over a ~2500-m-thick section of sedimentary units that are interbedded with and overlie CAMP lavas in the Hartford basin, which together represent the initial 2.4 Ma of the Jurassic according to cycle stratigraphic analysis. Characteristic directions carried by hematite were isolated by thermal demagnetization in 71 sites and define a coherent magnetostratigraphy supported by a positive reversal test and an interbasin fold test. Despite a pronounced overall normal polarity bias (only three relatively short reverse polarity intervals could be confirmed in the sampled section), normal polarity Chron H24n that encompasses the CAMP extrusive zone is no more than 1.6 Ma in duration. Elongation/inclination analysis of the 315 characteristic directions, which have a flattened distribution, produces a result in agreement with a published mean direction for the CAMP volcanic units as well as published results similarly corrected for inclination error from the Newark basin. The three data sets (CAMP volcanics, Newark corrected sediments, and Hartford corrected sediments) provide a 201 Ma reference pole for eastern North America at 67.0°N, 93.8°E, A_{95} = 3.2°. Paleopoles from the Moenave and Wingate formations from the Colorado Plateau that virtually define the J1 cusp can be brought into agreement with the 201 Ma reference pole with corrections for net clockwise rotation of the plateau relative to eastern North America and presumed sedimentary inclination error. The corrected data show that apparent polar wander for North America proceeds directly toward higher latitudes over the Late Triassic and Early Jurassic with no obvious change that can be associated with CAMP.

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1. Introduction

[2] The recent recognition of what may be the largest igneous province on Earth, the ~200 Ma central Atlantic magmatic province (CAMP) [*Marzoli et al.*, 1999] (Figure 1), with its close temporal proximity to major biotic turnover at the Triassic-Jurassic boundary [*Olsen*, 1999], adds impetus for seeking confirmation of possibly related phenomena. One is an apparently extended interval of pronounced normal polarity bias that has been found in several data compilations. An early global assessment of paleomagnetic data by [*Irving and Pullaiah*, 1976] suggested there was a poorly defined normal polarity interval in the Triassic, which roughly coincided with the Graham normal interval of *McElhinny and Burek* [1971] and the Newark normal interval of *Pechersky and Khramov* [1973]. Although no supporting evidence of an extended normal polarity interval has subsequently been found in magnetostratigraphic data for the Triassic [*Steiner et al.*, 1989; *Ogg and Steiner*, 1991; *Gallet et al.*, 1992; *Kent et al.*, 1995; *Muttoni et al.*, 1998; *Szurlies*, 2004], a ~10-Ma-long interval of pronounced normal polarity bias and low reversal frequency has been identified in the early Jurassic in several compilations of global paleomagnetic data [*Johnson et al.*, 1995; *Algeo*, 1996] and could conceivably be related to perturbation of the geodynamo by ascent of a mantle plume [e.g., *Larson and Olson*, 1991].[1]

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[1]Auxiliary materials are available at ftp://ftp.agu.org/apend/jb/2007jb005407.

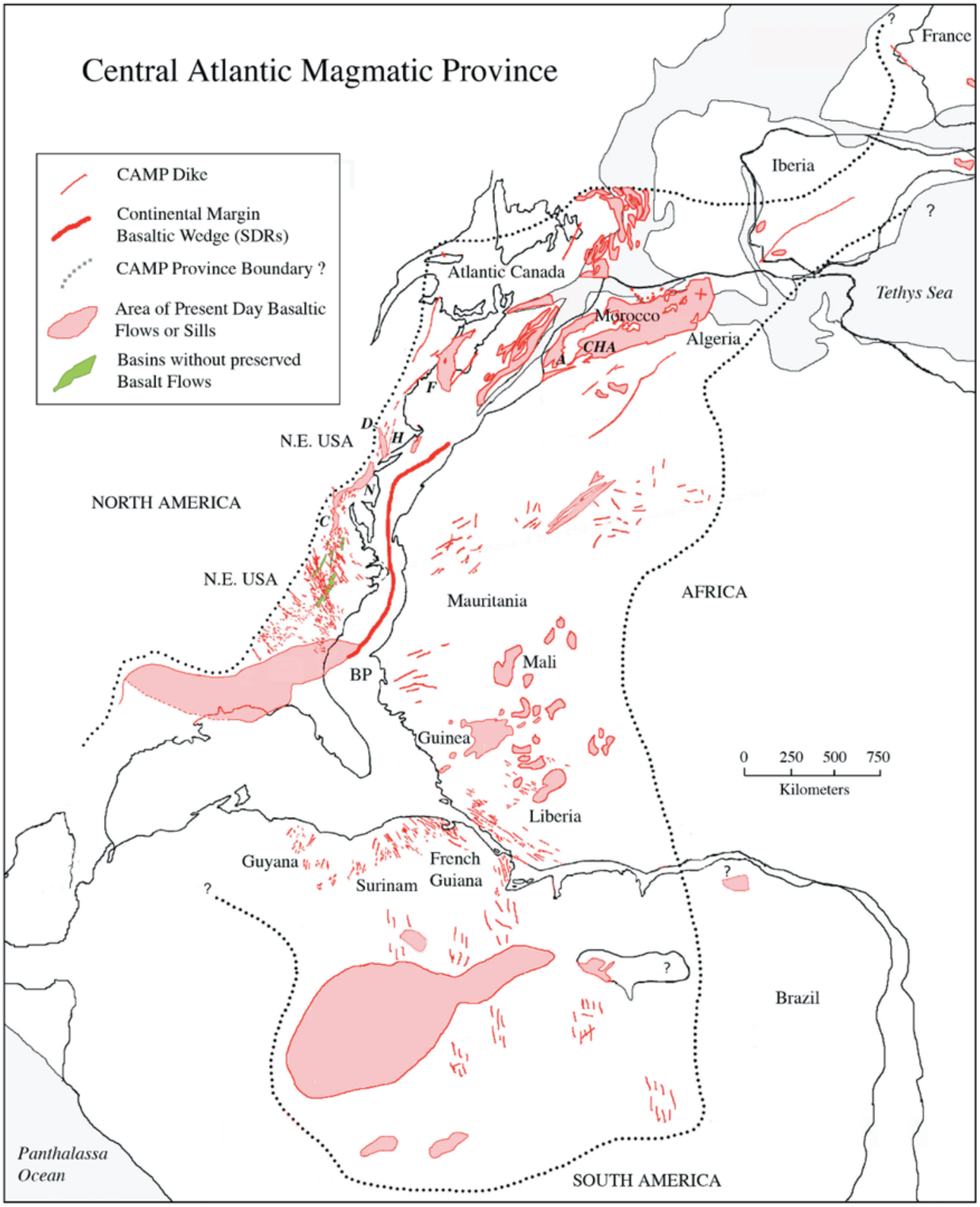

Figure 1. Paleogeographic extent of ~200 Ma central Atlantic magmatic province (CAMP) across the central Pangean supercontinent [after *McHone*, 2000; *Whiteside et al.*, 2007]. From south to north: BP, Blake Plateau; C, Culpeper basin; N, Newark basin; H, Hartford basin; D, Deerfield basin; F, Fundy basin; CHA, Central High Atlas basin; A, Argana basin.

[3] CAMP emplacement also seems to coincide temporally with an abrupt change in North American apparent polar wander (APW) at the so-called J1 cusp [*Gordon et al.*, 1984; *May and Butler*, 1986], which was suggested to reflect a major plate reorganization [*Gordon et al.*, 1984] or an episode of true polar wander [*Marcano et al.*, 1999], either of which could have been associated with the emplacement of CAMP. Late Triassic paleopoles from the Newark Supergroup are coherent between basins and do not have cusp-like directions [*Witte et al.*, 1991; *Kent et al.*, 1995; *Kent and Olsen*, 1997] whereas the apparent trend toward the J1 cusp in coeval paleopoles from the southwestern United States can be explained by a ~10–15° clockwise rotation of the Colorado Plateau region [*Kent and Witte*, 1993; *Steiner and Lucas*, 2000]. However, early Jurassic paleopoles from the Moenave and Wingate formations on the Colorado Plateau [*Ekstrand and Butler*, 1989; *Molina-Garza et al.*, 2003] that virtually define the J1 cusp

have few reliable counterparts from other parts of North America, where J1 cusp-like directions have thus far been reported only as overprints with uncertain age control from Texas [*Molina-Garza et al.*, 1995] and in some baked sediment sites in contact with CAMP-related igneous intrusions, and thus representing abbreviated recordings of the paleofield with uncertain structural control, in eastern North America [*Kodama et al.*, 1994].

[4] Although there is no obvious evidence of either a normal polarity superchron or the J1 cusp in the Late Triassic record from the Newark Supergroup, more data are clearly needed to document the polarity history and APW for North America in the early Jurassic when the CAMP event actually occurred. Paleomagnetic and chronostratigraphic data are already available from the Late Triassic and earliest Jurassic history of the Newark basin and provide some of the best available age constraints on the age and duration of CAMP igneous activity [*Olsen et al.*, 1996b; *Hames et al.*, 2000]. The oldest lavas in the Newark basin immediately postdate (within ~40 ka by cycle stratigraphy) the Triassic-Jurassic boundary identified on the basis of palynoflora and vertebrate (mainly footprint) evidence coinciding with a small iridium anomaly in a boundary clay layer [*Olsen et al.*, 2002a]. A reverse polarity interval (Chron E23r) occurs just below the Triassic-Jurassic boundary; with an estimated duration of only ~20,000 y, it is the shortest polarity interval among the 60 polarity chrons delineated in the astronomical polarity timescale based on the Newark succession [*Kent and Olsen*, 1999]. Despite its brevity, the occurrence of Chron E23r in close proximity to the Triassic-Jurassic boundary has evidently made it a beguiling target for correlation of distant sections, such St. Audrie's Bay in Britain [*Hounslow et al.*, 2004] and the High Atlas of Morocco [*Marzoli et al.*, 2004]. The succeeding normal polarity interval (Chron E24n) already encompasses more than 1000 m of CAMP lavas and interbedded sedimentary formations [*McIntosh et al.*, 1985; *Witte et al.*, 1991], making it the thickest polarity zone in the Newark basin stratigraphic succession [*Kent and Olsen*, 1999] even though its upper limit has not been found in the overlying Boonton Formation, which apart from becoming conglomeratic close to the border fault, is largely buried by Pleistocene glacial deposits. The full duration of Chron E24n, whose known record already constitutes one of the longer polarity intervals in the Newark geomagnetic polarity timescale (GPTS) [*Kent and Olsen*, 1999], has yet to be determined. Correspondingly, the search for J1 cusp directions has not extended over more than about the first ~600 ka of the Jurassic in the Newark Supergroup record.

[5] More extensive Jurassic age deposits are preserved in the nearby Hartford basin, another of the series of exhumed continental rift basins outcropping from Nova Scotia to North Carolina that are filled with continental strata of the Newark Supergroup (Figure 2a). Of all the exposed Mesozoic rift basins in eastern North America, the Hartford basin has the thickest section of continental strata of Early Jurassic age, totaling at least 4500 m (Figures 2b and 2c). These Jurassic age strata have long been recognized as containing red beds and cyclical lacustrine sediments [*Hubert et al.*, 1992], but there has been no comprehensive attempt to develop a magnetostratigraphy or to describe the cyclical sequence as a whole. In this paper we focus on the paleomagnetism and cyclostratigraphy of the lower 2500 m of the Jurassic age section, which is the fine grained and cyclical portion that begins with sedimentary units (Shuttle Meadow and East Berlin formations) that are interbedded with the CAMP lavas and extends into the lower half of the Portland Formation (Figure 2c). We describe and interpret the cyclicity in terms of Milankovitch orbital variations [*Olsen and Kent*, 1996], providing a basis for extending the astronomically calibrated GPTS for the Late Triassic [*Kent and Olsen*, 1999] into the Early Jurassic and a chronostratigraphic context for paleopoles.

2. Geologic Framework of Hartford Basin

[6] The Newark, Hartford and related early Mesozoic continental rift basins that are preserved on the margins of the Atlantic-bordering continents formed during the incipient breakup of Pangea in the Triassic (Figure 1). The overall structure of the Hartford basin (Figure 2b) is consistent with a step-faulted half graben geometry as seen in other Newark Supergroup basins [*Schlische*, 1993]. However, unlike the other exposed rift basins in eastern North America that have their long axes oriented northeast-southwest, the Hartford basin runs nearly north-south with a segmented west dipping border fault system on its eastern side toward which the basin strata tilt at predominately low to moderate (~10–15°) dips. The border fault system generally parallels the structural fabric of Paleozoic metamorphic basement, suggesting that the border faults may be reactivated structures [*Wise and Robinson*, 1982]. Numerous generally northeast trending intrabasinal faults with strike-slip and down-to-the-west normal offsets occur especially in the southern and central portions of the basin [e.g., *Davis*, 1898; *Sanders*, 1970]. These faults, as well as a series of transverse folds that increase in amplitude and frequency to the east toward the border fault system [*Wheeler*, 1937; *Schlische*, 1995], complicate the homoclinal geometry of the basin strata. Fission track analyses suggest moderate (2–5 km) burial depths [*Roden and Miller*, 1991; *Roden-Tice and Wintsch*, 2002].

[7] The large-scale lithostratigraphy of the basin fill is composed of a tripartite succession (Figure 2c): (1) a lower coarse arkosic unit up to 3000 m thick named the New Haven Formation; (2) a middle, generally finer grained sequence (the focus of this paper) containing interbedded basalt flows of the CAMP near its base, which together are about 2500 m thick and, consist of, in ascending order, the Talcott Basalt, Shuttle Meadow Formation, Holyoke Basalt, East Berlin Formation, Hampden Basalt, and about the lower half of the Portland Formation; and (3) an upper coarse arkosic unit that exceeds 1500 m in thickness comprising the upper Portland Formation [e.g., *Krynine*, 1950; *Sanders*, 1968]. A Jurassic age for the middle and upper succession is based on both palynology [*Cornet et al.*, 1973; *Cornet and Traverse*, 1975; *Cornet and Olsen*, 1985] and vertebrate biostratigraphy [*Lucas and Huber*, 2003; *Olsen and Galton*, 1977]. Radioisotopic dates from the CAMP basaltic flows interbedded with these strata have substantial scatter attributed in large part to postcooling alteration but are not inconsistent with a Jurassic age [*Seidemann*, 1989]. Geochemical and cyclostratigraphic correlation with other basins that have igneous rocks with

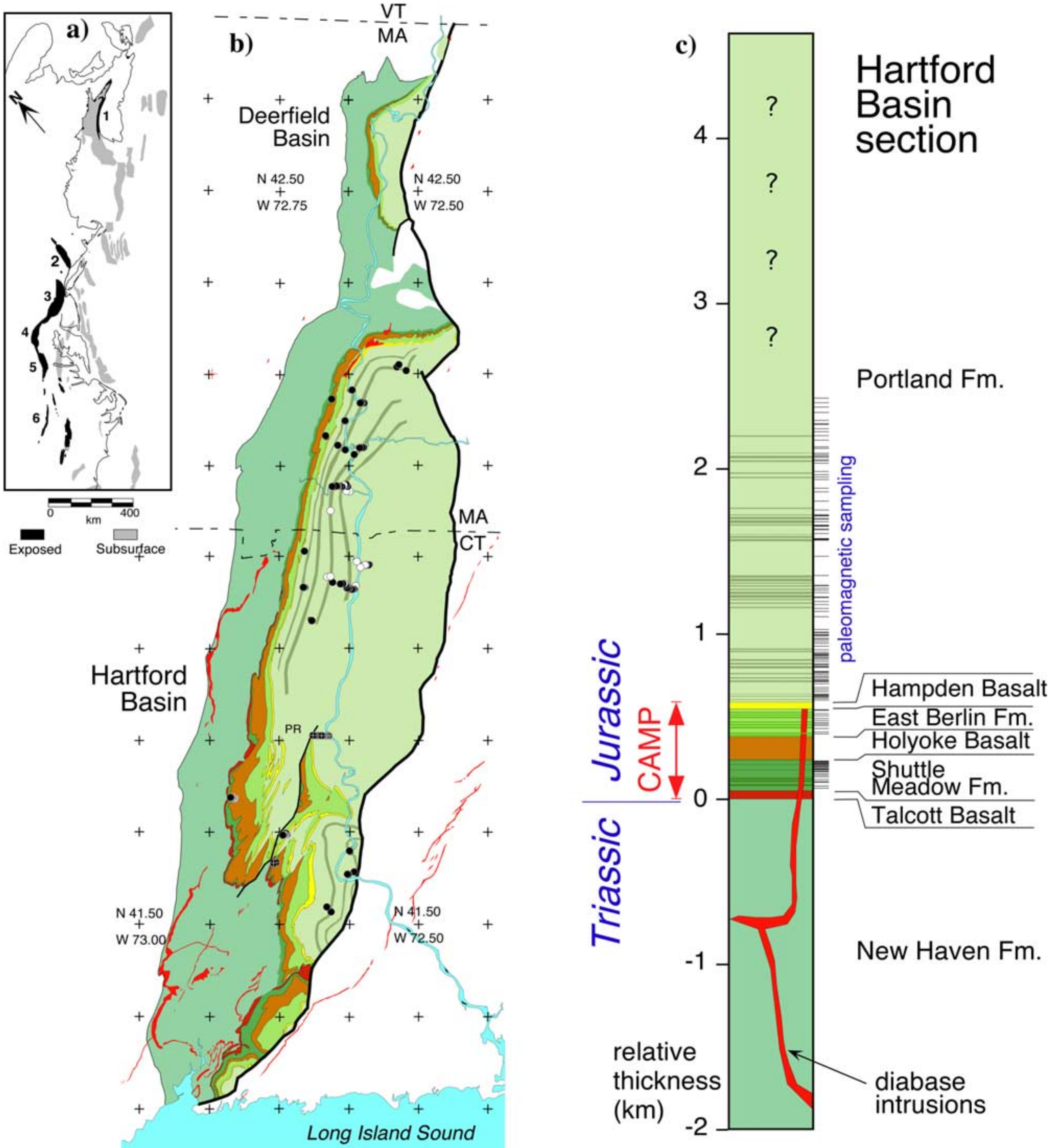

Figure 2. (a) Early Mesozoic rift basins in eastern North America: 1, Fundy; 2, Hartford; 3, Newark; 4, Gettysburg; 5, Culpeper; 6, Danville. (b) Geologic sketch map of Hartford basin with sampling sites as open (reverse polarity) and filled (normal and indeterminate polarity) circles. Army Corps of Engineers Park River drainage project geotechnical cores are indicated by a series of filled circles with crosses next to label "PR". (c) Stratigraphic section of Newark Supergroup in Hartford basin with tics along right margin of column showing paleomagnetic sampling levels in Shuttle Meadow, East Berlin, and lower Portland formations.

more secure radioisotopic dates [*Sutter*, 1988; *Dunning and Hodych*, 1990; *Hames et al.*, 2000] support an earliest Jurassic (~200 Ma) age for the basalts [*Olsen et al.*, 2003]. Mafic igneous sills (e.g., Barndoor, West Rock, East Rock, Carmel) and dikes (e.g., Buttress, Higganum) intrude the New Haven Formation; some of the dikes evidently served as feeders for the tholeiitic lava flows [*Philpotts and Martello*, 1986] but they do not seem to cut the Portland

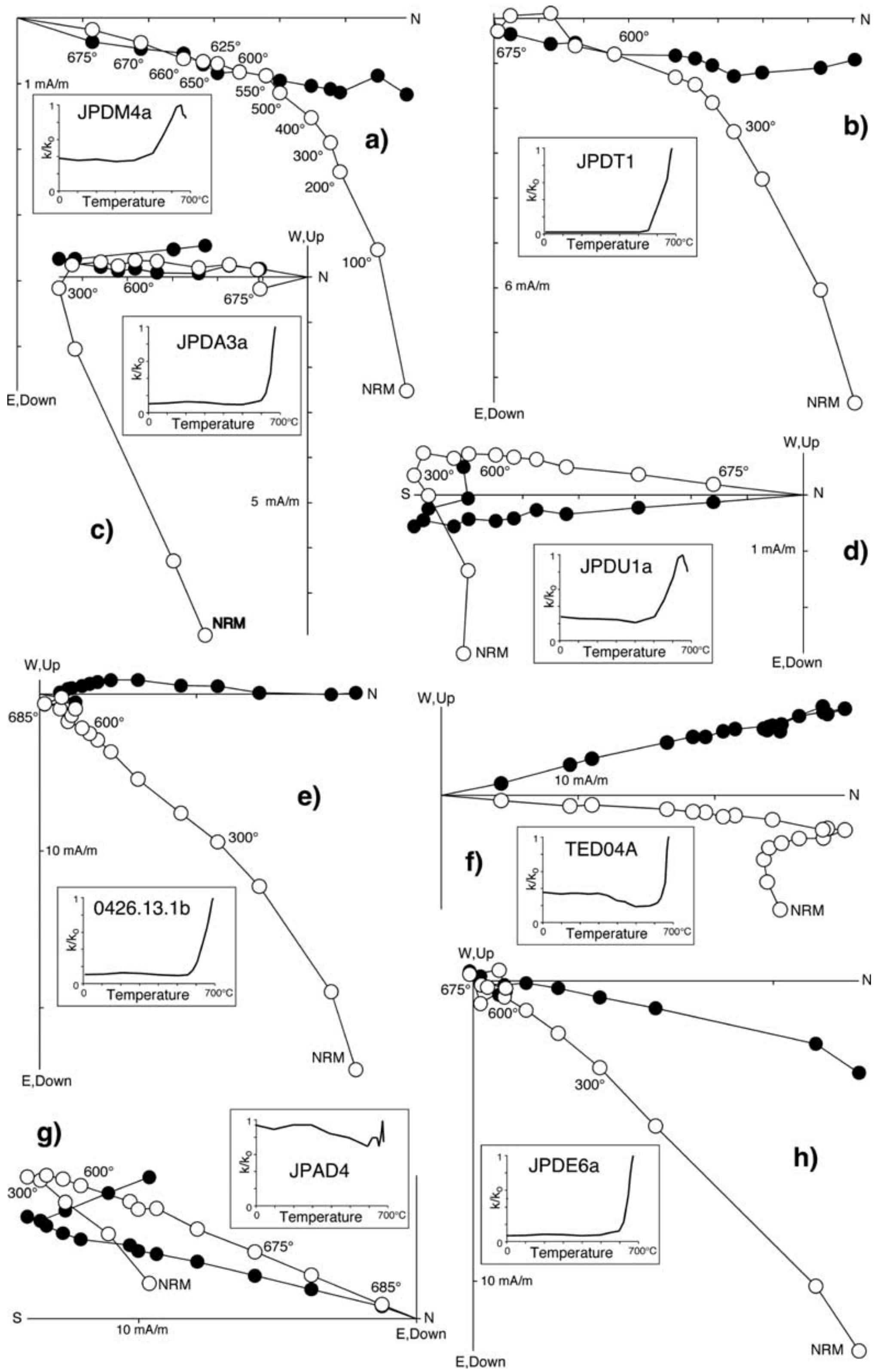

Figure 3. Representative vector end-point diagrams (open/filled circles are projections on vertical/horizontal axes in geographic coordinates with temperatures in centigrade adjacent to selected points) of thermal demagnetization of NRM of samples from Portland (Figures 3a–3d, 3g, and 3h), Shuttle Meadow (Figure 3e), and East Berlin (Figure 3f) formations. These samples were selected from magnetozones (b, e, f) H24n, (d) H24r, (h) H25n, (c) H25r, (a) H26n, and (g) H26r (see Figure 8 for magnetostratigraphy). Insets show changes in magnetic susceptibility (k/k_0) normalized to peak value after each heating step as a monitor of magnetochemical alteration during thermal demagnetization.

Formation. Parenthetically, the Deerfield basin is connected to the Hartford basin (Figure 2b) and has an analogous stratigraphic development except that only one basalt unit (Deerfield Basalt, equivalent to the Holyoke Basalt of the Hartford basin [*Luttrell*, 1989; *Prevot and McWilliams*, 1989; *Tollo and Gottfried*, 1992]) is present.

[8] Published paleomagnetic work in the Hartford basin has focused almost entirely on the CAMP lavas and dikes [*DuBois et al.*, 1957; *Irving and Banks*, 1961; *De Boer*, 1968; *Prevot and McWilliams*, 1989; *Smith*, 1976; *Smith and Noltimier*, 1979] (see summary by *Hozik* [1992]). A notable exception is the early work of *DuBois* [1957], who reported paleomagnetic results from 32 samples of "Triassic rocks from the Connecticut Valley in the State of Connecticut", whose sampling localities are otherwise not described. Normal and reverse polarity directions were reported by *DuBois* [1957], but to our knowledge, there has been no follow-up paleomagnetic work on sediments of the Hartford basin until the present study. From the contiguous Deerfield basin, paleomagnetic results from two sedimentary sites (from the Fall River Formation, a unit correlative to the Shuttle Meadow Formation of the Hartford basin) were reported by *McEnroe and Brown* [2000].

3. Paleomagnetic Sampling and Measurements

[9] The Jurassic sedimentary units in the Hartford basin (Shuttle Meadow, East Berlin, and Portland formations) were the focus of sampling. Samples were collected from 80 sites in stream and road cuts with a gasoline-powered portable drill and oriented with a magnetic compass. A sampling site typically consisted of four to six oriented cores covering several meters of section; sampling at three sites ranged over several tens of meters of section and included 8 to 12 samples each, whereas sampling at five sites included only a single oriented hand sample from which up to three specimens were cut and measured. Bedding attitude was measured at every sampling site for tilt corrections. Outcrops were sporadic due to the low topographic relief and shallow bedding dips; hence, a stratigraphic composite section was assembled from a number of across-strike profiles that were linked by tracing and mapping distinctive beds over the course of numerous field trips to the area. Most of the sampling sites in the Portland Formation were collected in two traverses, the Stony Brook (Connecticut) and Westfield River (Massachusetts) sections; additional stratigraphic coverage was obtained from around Holyoke and South Hadley Falls (Massachusetts) along the Connecticut River, the Chicopee River around Chicopee (Massachusetts) and in the Durham-Portland area in Connecticut. Geologic mapping in the Portland Formation was required to link the sampling sites into a composite section and resulted in the recognition of new lithostratigraphic members that also provide a cycle stratigraphic framework [*Olsen et al.*, 2005]. We believe that our sampling at the 80 sites virtually exhausts available outcrop of the Shuttle Meadow, East Berlin and lower Portland formations. However, we were able to fill in some gaps for magnetostratigraphy and description of lithostratigraphy for the poorly exposed lowest portion of the Portland Formation by gaining access to some relatively short engineering cores for the Park River project, as described below.

[10] All magnetic remanence measurements were made on a 2G three-axis DC-SQUID superconducting rock magnetometer housed in a magnetically shielded room (<1000 nT). Thermal demagnetization was performed in a custom-built oven with three independent heating zones and a water jacket for reproducible temperature control and housed in high magnetic permeability shields for a low magnetic field environment (<5 nT) that is critical for resolving ancient magnetizations that can be masked by spurious magnetizations introduced by lab-induced thermochemical alteration, which was monitored by measuring the magnetic susceptibility with a Bartington instrument after each thermal demagnetization step.

[11] Progressive thermal demagnetization of NRM reveals a relatively straightforward two-component structure in most samples, comprising a low unblocking temperature (up to 300°C) A component that tends to be aligned about the northerly and down present-day field direction followed by a characteristic magnetization (ChRM) with shallow directions to the north (Figure 3a, 3b, 3e, 3f, and 3h) or to the south (Figure 3c, 3d, and 3g) that converges toward the origin by maximum unblocking temperatures of 685°C. The only systematic departure from this pattern was observed in samples from a site in the East Berlin Formation, which tended to show back-tracking in demagnetization trajectories between about 200° and 400°C (Figure 3f) that could represent a partial reverse polarity overprint embedded between normal polarity A and ChRM components. What is absent in the Hartford samples, however, is a distinct northerly and moderately down component with intermediate unblocking temperatures ($\sim$300° to at least 600°C) that was found to be ubiquitous in sedimentary rocks in the Newark and Dan River basins and attributed to a Middle Jurassic remagnetization event [*Witte and Kent*, 1991; *Kent et al.*, 1995; *Kent and Olsen*, 1997]. It is possible that the Hartford basin escaped this remagnetization event, perhaps because of its shallower burial and/or lower thermal maturation [*Pratt et al.*, 1988; *Roden and Miller*, 1991]. Alternatively, the remagnetization event as identified in the Newark and Dan River basins had in fact occurred during CAMP igneous activity and is therefore not expected to be as pronounced or ubiquitous in the Hartford basin rocks we sampled, which are of CAMP age or younger.

4. IRM Analysis

[12] The dominant magnetic mineralogy was characterized using isothermal remanent magnetization (IRM) acquisition to 2.5 T using an ASC pulse magnetizer and thermal demagnetization of orthogonal axes IRM by the method of *Lowrie* [1990]. Most of the sampling was done with a preference for red mudstones and siltstones rather than the gray shales and thus a hematite carrier of remanence is expected in most of the sites. IRM acquisition curves (Figure 4a) generally show a gradual approach to saturation to 2.5T although the gray samples (e.g., JPDA6b, JPDE6b) have remanent coercivities of around 170 mT, which are considerably lower than around 600 mT for the red samples (e.g., JPDM4b, JPDU1b, 160702B05). Thermal demagne-

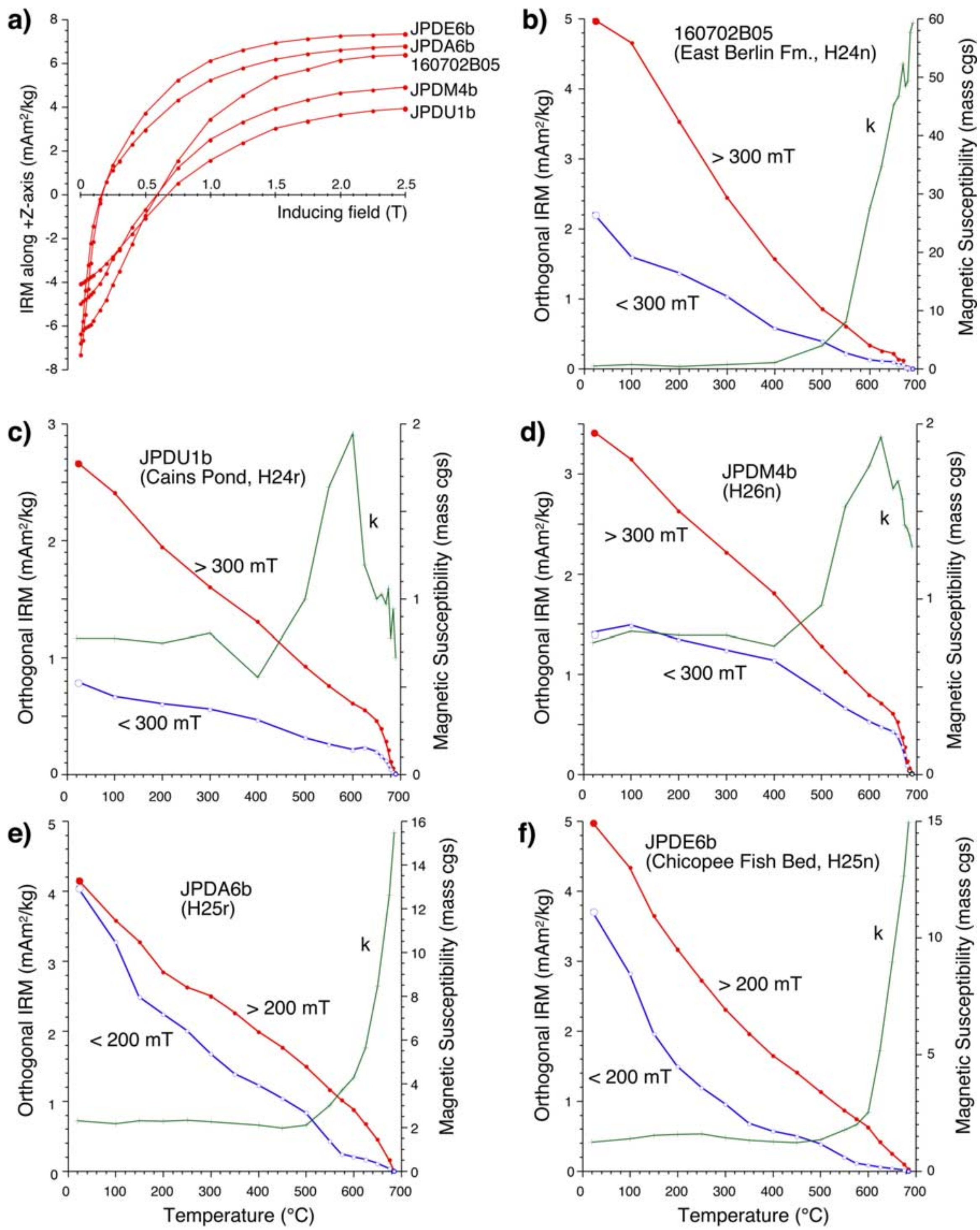

Figure 4. (a) IRM acquisition and (b, c, d, e, f) thermal demagnetization of orthogonal IRM back-field components of lower (<200 mT or <300 mT) and higher (>200 mT or >300 mT) coercivity with attendant changes in magnetic susceptibility (k) of representative sedimentary rock samples from the Hartford basin. In this method of IRM acquisition (Figure 4a), remanent coercivity is the intersection of the acquisition curve and null IRM. Samples come from red beds (Figure 4b for East Berlin Formation; Figures 4c and 4d for Portland Formation) or from gray shales (Figures 4e and 4f for Portland Formation) and represent sites with either normal polarity (Figures 4b, 4d, and 4f) or reverse polarity (Figures 4c and 4e) characteristic directions (see NRM thermal demagnetization data from some companion specimens in Figure 3).

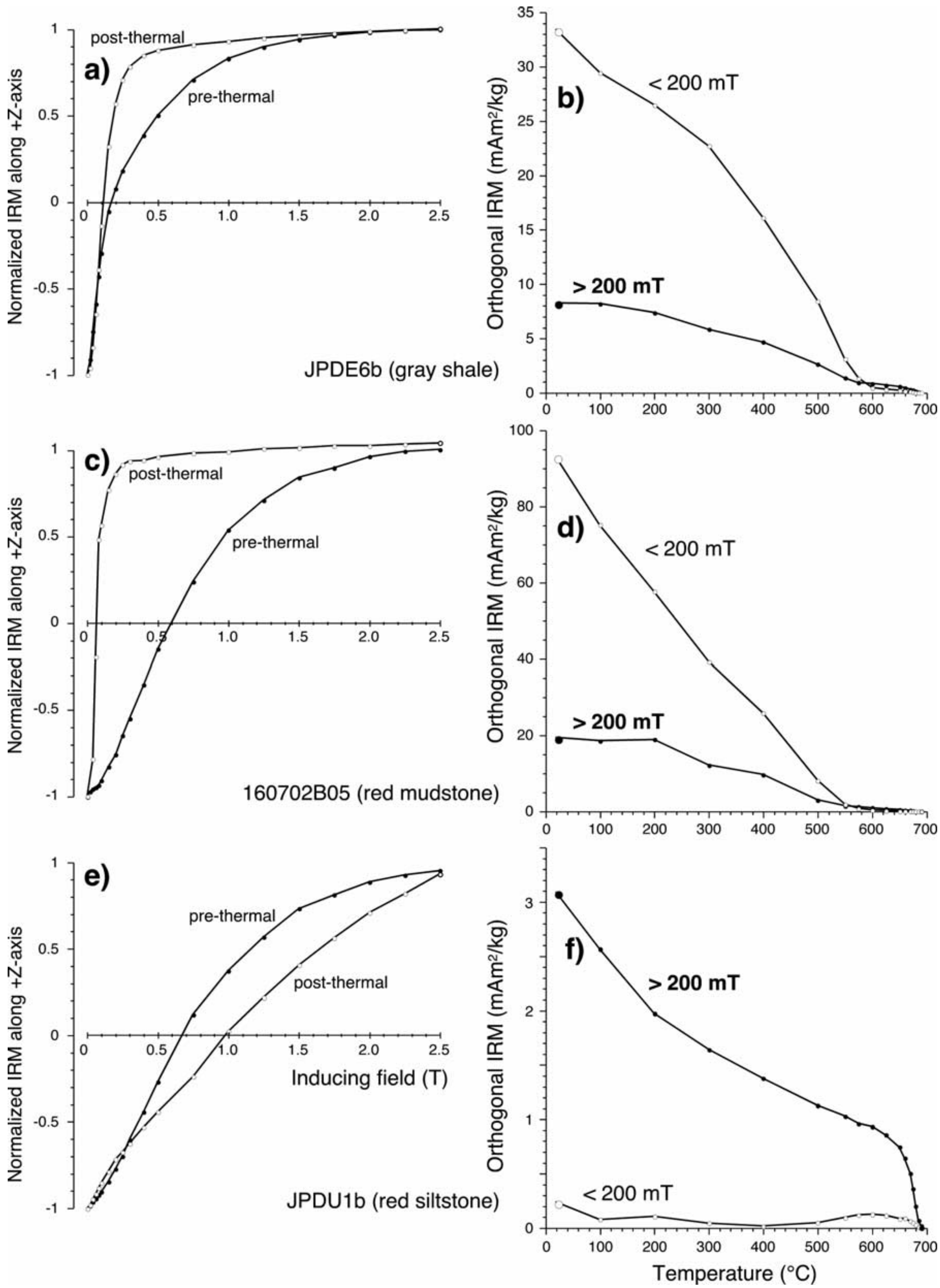

Figure 5. (left) Comparison of IRM back-field acquisition before and after the samples were thermally demagnetized to 685°C. In this method, remanent coercivity is the intersection of the acquisition curve and null normalized IRM. (right) Thermal demagnetization of orthogonal IRM given after heating to 685°C can be compared to initial thermal demagnetization of IRM of the same samples in Figure 4. Samples are from (a, b) gray shale from Portland Formation, (c, d) red mudstone from East Berlin Formation, and (e, f) red siltstone from Portland Formation.

tization of IRM (Figures 4b–4f) confirms that the dominant and higher coercivity (>200 or >300 mT) component of an orthogonal IRM is invariably associated with a maximum unblocking temperature of about 685°C, which is compatible with hematite. The lower coercivity (<200 or <300 mT) component of the orthogonal IRM shows an inflection in its thermal demagnetization spectrum at around 580°C suggesting the presence of magnetite in the gray samples (Figures 4e and 4f), whereas the red samples are dominated by hematite (maximum unblocking temperature of about 685°C) in both high-and low-coercivity fractions.

[13] Hematite-dominated sampling sites can posses either normal or reverse polarity (northerly or southerly) characteristic directions, for example, sample JPDU1b is from a reverse polarity site (Figure 4c) and samples 160702B05 and JPDM4b are from normal polarity sites (Figures 4a and 4d); likewise for sampling sites in gray beds that show contributions from magnetite, for example, sample JPDA6b is from a reverse polarity site and sample JPDE6b is from a normal polarity site (Figures 4e and 4f). Nevertheless, the characteristic remanence is invariably associated with the hematite phase (unblocking temperature above ~600°C) even in the gray beds that also have a magnetite contribution (Figures 3c and 3h). We suggest that the characteristic remanence is carried by detrital hematite or an early authigenic product that was acquired during or soon after deposition whereas the sporadic presence of magnetite that is mostly restricted to gray shales, such as the Chicopee fish bed, may reflect its preservation or production in a localized reducing depositional environment.

[14] Magnetochemical alteration of the samples during thermal demagnetization is common as indicated by changes in magnetic susceptibility, which usually starts to increase noticeably after about the 500°C step. In the gray samples, magnetic susceptibility often continues to increase by more than an order of magnitude by 680°C (Figures 4e and 4f). The large monotonic susceptibility rise with thermal treatment is associated with the production of magnetite, as revealed by IRM acquisition (Figure 5a) and thermal demagnetization of orthogonal IRM experiments on previously heated samples (Figure 5b), showing that a large IRM phase that approaches saturation by 300 mT and has maximum unblocking temperatures around 575°C becomes the dominant magnetization component after heating. In contrast, red siltstone samples that had only modest susceptibility changes with thermal treatment (e.g., Figure 4c) maintained a predominantly hematite mineralogy characterized by lack of saturation and high (~685°C) maximum unblocking temperatures (Figures 5e and 5f). Nevertheless, sediment grain size rather than just color (or initial magnetic mineralogy) seems to be an important determinative factor in the thermal alteration profile. Gray samples that showed large susceptibility changes with heating were typically fine-grained mudstones or shales; however, some mudstones are also reddish and although their magnetizations are dominated by hematite, many of these samples showed large susceptibility increases with initial heating (e.g., sample 160702B05; Figure 4a) that were also associated with the production of magnetite (Figures 5c and 5d), as in the gray shales. We suspect that the dramatic magnetochemical alteration seen in laboratory heating of the fine-grained gray shales and red mudstones is probably due to the breakdown of clays. Clays are much less abundant in the red siltstones and fine-grained sandstones, making such hematite-bearing rocks less prone to magnetochemical alteration and thereby enhancing their suitability for paleomagnetic study.

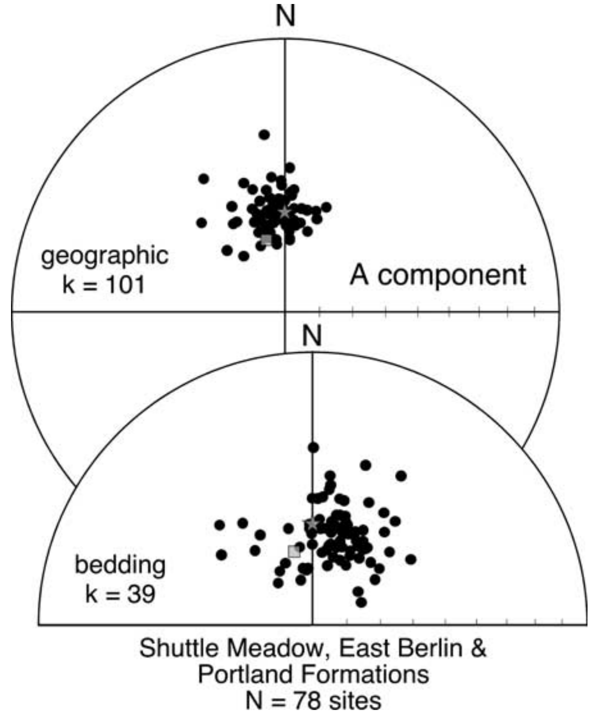

Figure 6. A component site-mean directions (filled circles) from Shuttle Meadow, East Berlin, and lower Portland formations before (geographic) and after (bedding) correction for tectonic bedding tilts, plotted on lower hemisphere of equal-area plots; star is geocentric axial dipole field direction, and square is present-day direction. The decrease in precision parameter (k) after tilt correction is significant and indicates a negative fold test.

5. NRM Directions

[15] NRM component vectors were estimated using principal component analysis [*Kirschvink*, 1980] on demagnetization trajectories typically from 100° to 300°C for the A component and anchored to the origin from 600° to 675°C for the ChRM (except for three sites in fine-grained gray to purplish shales (JPAA and JPAB in the lowermost Portland Formation, and 160702B in the East Berlin Formation) that were analyzed for ChRM between 400°C and 600°C due to the onset of erratic directions associated with large susceptibility increases that occurred at higher temperatures). Line fits with MAD angles greater than 18° were rejected as were data from a few samples that were obviously misoriented when compared with other samples at a site. Out of a total of 398 samples measured and analyzed, over 83% (331 from 78 sites) yielded an acceptable A component direction (median and mean MAD angles of 2.8° and 3.8°, respectively) and nearly 80% (315 from 71 sites) yielded acceptable estimates of a ChRM direction (median and mean MAD angles of 3.5° and 5°, respectively).

Table 1. Paleomagnetic Directions Isolated From Early Jurassic Sedimentary Rock Units From the Hartford Basin[a]

Rock Unit	N	Geographic Dec	Geographic Inc	Geographic k	Geographic α_{95}	Tilt Corrected Dec	Tilt Corrected Inc	Tilt Corrected k	Tilt Corrected α_{95}
A Component (100–300°C)									
SM+EB+PF	78	353.9°	60.2°	101	1.6°	13.4°	62.3°	39	2.6°
ChRM (600–675°C)									
SM	8	356.4	20.4	60	7.2	1.2	20.4	64	7.0
EB	5	358.3	21.6	63	9.7	7.7	28.8	41	12.2
SM+EB	13	357.1	20.9	66	5.1	3.6	23.6	47	6.1
PF									
Normal	45	5.4	20.2	25	4.3	9.2	22.5	23	4.5
Reverse	13	178.5	−17.1	12	12.6	181.6	−19.6	12	12.5
All	58	3.8	19.6	20	4.3	7.5	21.9	19	4.4
SM+EB+PF									
Normal	58	3.5	20.4	28	3.6	7.9	22.7	26	3.7
Reverse	13	178.5	−17.1	12	12.6	181.6	−19.6	12	12.5
All	71	2.6	19.8	23	3.6	6.8	22.2	21	3.7

[a]Rock units are SM, Shuttle Meadow; EB, East Berlin; PF, Portland Formation. N is number of sites, Dec and Inc are the declination and inclination, k is the best estimate of Fisher's precision parameter, and α_{95} is the radius of 95% circle of confidence for the mean direction.

[16] The 78 site-mean A component directions from the Shuttle Meadow, East Berlin and Portland formations are well grouped in geographic coordinates about a mean direction of declination (D) = 353.9°, inclination (I) = 60.2°, with a radius of the circle of 95% confidence (α_{95}) = 1.6° (Figure 6). This is close to a modern field direction (D = 346° I = 68° for present-day field, D = 0° I = 61° for geocentric axial dipole field); moreover, the distribution is significantly more dispersed (precision parameter k decreasing by a factor of 2.6; Table 1) after bedding tilt corrections, indicative of a negative fold test. The low unblocking temperature A component is thus most probably a recently acquired viscous magnetization.

[17] The 13 fully oriented sites from the Shuttle Meadow and East Berlin formations all have northerly and shallow (normal polarity) ChRM directions, whereas in the case of the 58 sites from the Portland Formation, 45 had northerly and shallow (normal polarity) ChRM directions and 13 sites had southerly and shallow (reverse polarity) ChRM directions (Figure 7). The combined tilt-corrected mean normal (58 sites) and reverse (13 sites) directions depart by only 6.7° from antipodal, which is less than the critical angle [*McFadden and McElhinny*, 1990], indicating a positive reversal test.

[18] The directional dispersion for the ChRM site means in the Shuttle Meadow, East Berlin and Portland formations hardly changes with corrections for bedding tilts because unlike the much steeper A component, the shallow ChRM directions tend to be closer to coaxial with the bedding strikes and thus less sensitive to dip corrections. However, there is sufficient variation in bedding attitudes with respect to available data from correlative sedimentary rock units in the Newark basin (Feltville and Towaca formations [*Witte and Kent*, 1990]) for an interbasin fold test. The mean ChRM direction for the 71 tilt-corrected sites from the Shuttle Meadow, East Berlin and Portland formations (D = 6.8°, I = 22.2° α_{95} = 3.7°; Table 1) corresponds to a paleopole (59.0°N 94.5°E) for a nominal locality (42°N

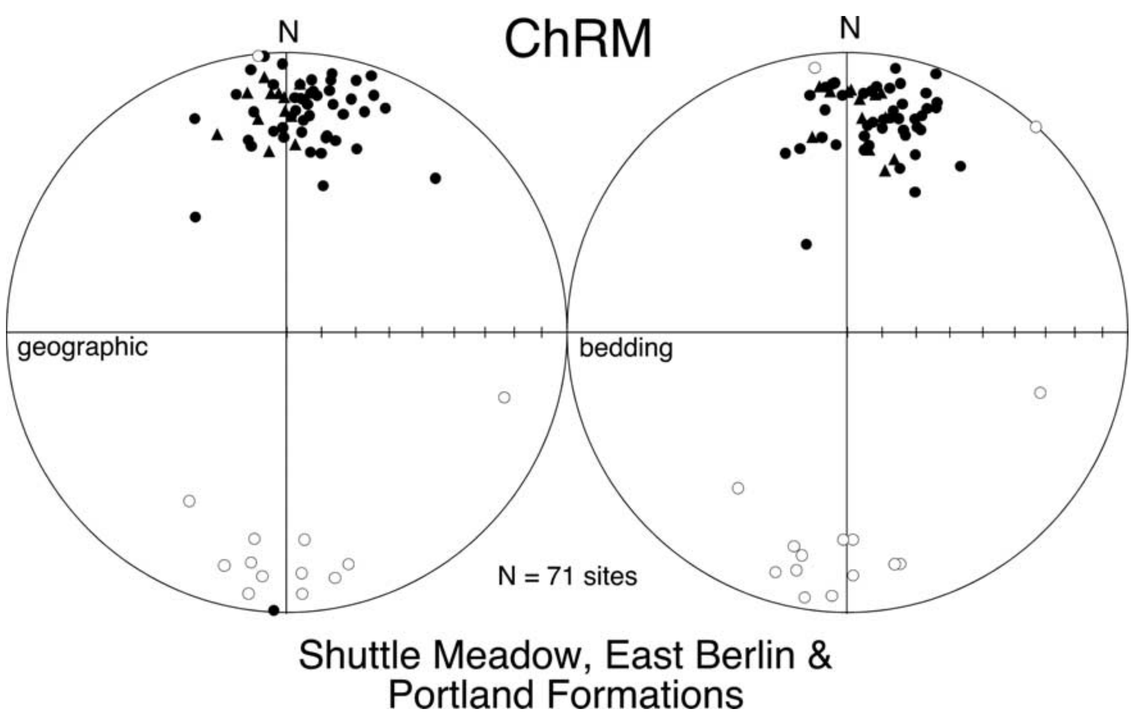

Figure 7. Characteristic (ChRM) site-mean directions from Shuttle Meadow and East Berlin formations (triangles) and from lower Portland Formation (circles). Open/filled symbols are plotted on upper/lower hemisphere of equal-area plots before (left, geographic) and after (right, bedding) correction for tectonic bedding tilts.

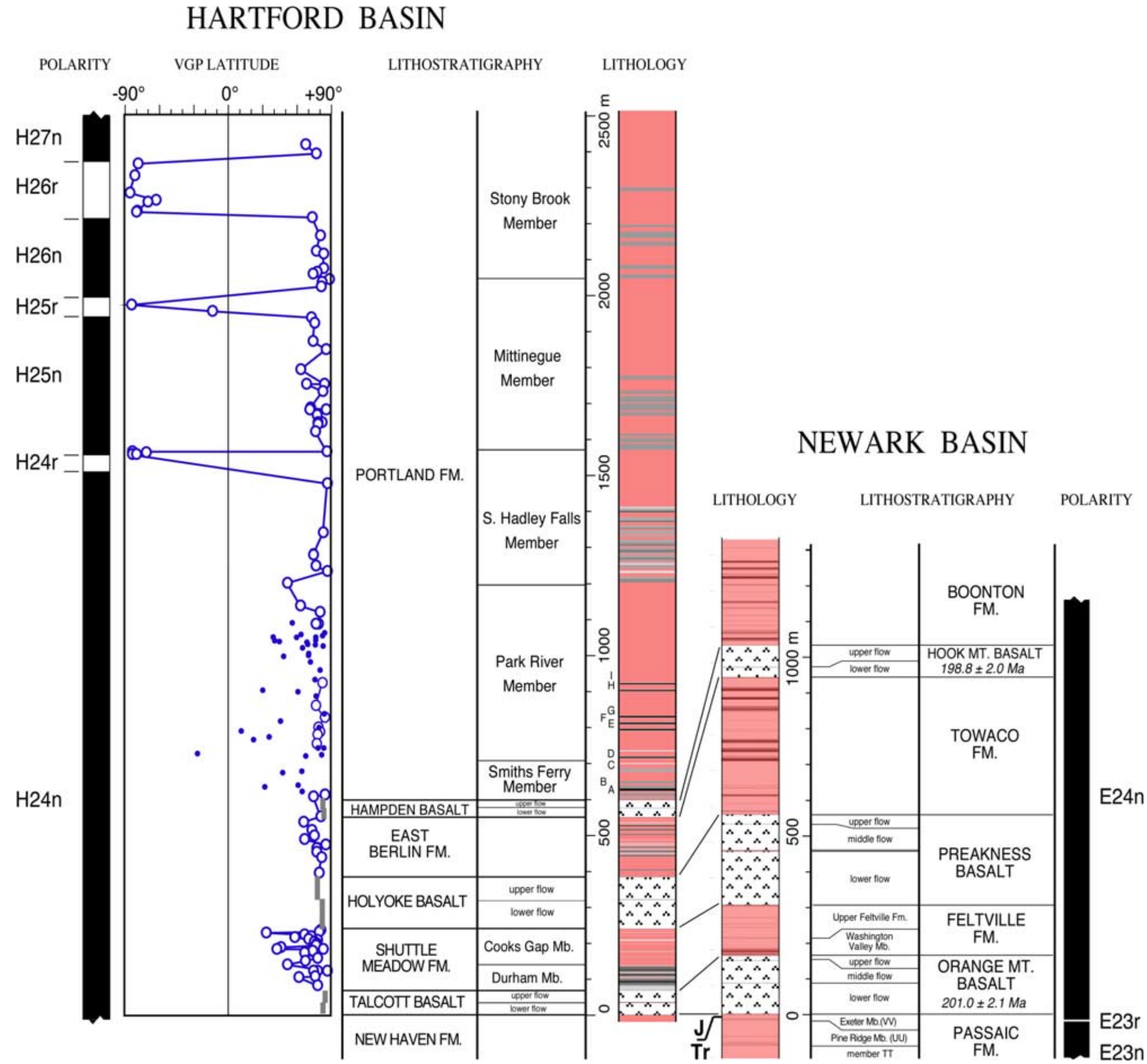

Figure 8. Magnetostratigraphy and lithostratigraphy of Early Jurassic strata in the Hartford basin. Magnetic polarity chrons are identified next to the polarity column where filled and open bars denote normal and reverse polarity interpreted from VGP latitudes (values approaching +90° indicate normal polarity and values approaching −90° indicate reverse polarity) with respect to overall mean paleopole for outcrop sites (open circles), sediment samples from Park River cores (small filled circles), and from lava units (bars [*Prevot and McWilliams*, 1989]). Letters adjacent to lithology column correspond to key beds used for correlation of Park River cores (see Figure 9). Correlative section from Newark basin is shown at right (lithology after *Olsen et al.* [1996b], polarity column from *Kent and Olsen* [1999]) with numerical ages for Orange Mountain and Hook Mountain basalts shown in lithostratigraphy panel from *Hames et al.* [2000].

72.5°W) in the Hartford basin that differs by only an insignificant 3.7° from the paleopole (55.3°N 94.5°E A_{95} = 5.4°) for 11 tilt-corrected sites in the Feltville and Towaco formations from the Newark basin extrusive zone reported by [*Witte and Kent*, 1990]. In geographic coordinates, however, the mean ChRM direction for the 71 sites from the Hartford basin (D = 2.6° I = 19.8° α_{95} = 3.6°; Table 1) gives a paleopole (58.1°N 102.7°E) that differs by a significant 7.8° from the Feltville and Towaco formations (56.1°N 90.0°E A_{95} = 5.7°) calculated without tilt correction [*Witte and Kent*, 1990]. The interbasin fold test is thus positive, which is also the case when ChRM directions from only the more strictly equivalent units (Shuttle Meadow and East Berlin formations versus Feltville and Towaco formations) are compared. The positive reversal test and positive fold test indicate that the ChRM of the Shuttle Meadow, East Berlin and Portland formations was acquired early in the history of the rock units and is relatively uncontaminated by secondary components.

6. Magnetostratigraphy

[19] For diagnosing the geomagnetic polarity of the characteristic magnetizations, a virtual geomagnetic pole (VGP) was calculated for each site direction and its latitude was compared to the north paleopole (59.0°N 94.5°E) corresponding to the overall ChRM mean direction. VGP latitudes close to +90° (or -90°) are interpreted as recording normal (or reverse) polarity of the early Jurassic geomagnetic field and are plotted with respect to stratigraphic level to develop a magnetostratigraphy.

[20] Previous paleomagnetic work has established that the lava units in the Hartford basin (Talcott, Holyoke, and Hampden basalts) erupted during normal geomagnetic polarity [e.g., *Prevot and McWilliams*, 1989]. This conclusion can now be extended to the entire CAMP extrusive zone of the Hartford basin because the interbedded sedimentary units, Shuttle Meadow and East Berlin formations, are also characterized by normal polarity (Figure 8). We cannot, of course, exclude the possibility of undetected short reverse polarity intervals but they would have to be less than about 20 m thick, or constitute no more than a few percent of the total stratigraphic thickness of the sedimentary units, at the present sampling density.

[21] The normal polarity interval in the Hartford extrusive zone encompasses approximately 600 m of section and most probably corresponds to the normal polarity interval (Chron E24n) of the homotaxial extrusive zone in the Newark basin, where the three lava units (Orange Mountain, Preakness and Hook Mountain basalts) as well as the interbedded sedimentary units (Feltville and Towaco formations), together about 1000 m thick, are also characterized by normal polarity [*McIntosh et al.*, 1985; *Prevot and McWilliams*, 1989; *Witte and Kent*, 1990; *Kent et al.*, 1995; *Kent and Olsen*, 1999] (Figure 8). For convenience and to emphasize this correlation, we refer to the Hartford extrusive zone normal polarity interval as magnetozone H24n. Suitable exposures of the New Haven Formation immediately below the Talcott Basalt could not be found to establish the expected presence of the thin reverse polarity interval corresponding to Chron E23r that occurs in the uppermost Passaic Formation and within a few meters below the Orange Mountain Basalt in the Newark basin [*Kent et al.*, 1995; *Kent and Olsen*, 1999].

[22] The outcrop sites from approximately the lower 950 m of the Portland Formation immediately overlying the youngest lavas (Hampden Basalt) are also characterized by normal polarity that can be regarded as an extension of magnetozone H24n (Figure 8). There are, however, several sampling gaps of 100 m or more, mainly because of cover by sediments of Pleistocene Lake Hitchcock in northern Connecticut and Massachusetts. We were able to fill several of these sampling gaps by gaining access to a series of geotechnical cores taken by the Army Corps of Engineers for the construction of the Park River drainage project in and near the city of Hartford (Figure 2b). The cores are presently stored at the Connecticut Department of Environmental Protection Western District Headquarters in Harwinton, Connecticut. A composite section of approximately 400 m thickness above the Hampden Basalt was assembled by projecting bedding attitudes and tracing a series of distinctive beds (informally labeled A to I) in a transect of 14 selected drill cores 10 cm diameter and up to 70 m long (Figure 9). Sample plugs were drilled perpendicular to the core axis (assumed to be vertical) and in the direction of bedding dip, which was assumed to be in the regional direction of east-southeast (120°) and used for azimuthal reorientation of the core segments. We analyzed 57 samples (up to 14 from each core) using the same laboratory measurement and thermal demagnetization techniques as for the outcrop samples. Well-defined, shallow ChRM directions were isolated from 600° to 675°C in 48 of the samples, although 9 of these samples had upward inclined A components, suggesting that these particular core segments had been inadvertently turned upside down during handling. VGP latitudes for the ChRM after azimuthal reorientation and correction for bedding tilt of the accepted 39 samples show appreciable scatter (Figure 9), which is not unexpected given the uncertainties in reorienting the samples using core-bedding plane intersections with shallow bedding dips. Nevertheless, the sample VGP latitudes are consistent with normal polarities, with one possible exception: a sample from near marker bed D, about 100 m above the Hampden Basalt (core FD-12T sample footage 198), had a negative albeit low VGP latitude, whose significance as an indication of a thin interval of reverse polarity, a polarity excursion, or simply noisy data is unclear and requires confirmation.

[23] In the outcrop sites, the first unambiguous reverse-polarity magnetizations are encountered near the top of the South Hadley Falls Member, approximately 950 m above the Hampden Basalt, where four closely spaced sites at Cains Pond record high southerly VGP latitudes (Figure 8). These reverse polarity sites cover an interval less than 10 m thick although available bounding constraints allow the reverse polarity interval (magnetozone H24r) to be as much as ~100 m thick. This reverse polarity magnetozone has no counterpart in the Newark basin, where the available section ends in normal polarity Chron E24n (= H24n).

[24] Another reverse-polarity magnetozone (H25r) of perhaps comparable (~100 m) thickness occurs above an intervening ~300 m interval of normal polarity of magnetozone H25n (Figure 8). Magnetozone H25r, which is in the uppermost Mittinegue Member, is delineated by only two sites, which are, however, located more than 30 km apart along-strike in the Agawam and Stony Brook sections. Within the middle and upper Stony Brook Member, a ~200 m interval of normal polarity (magnetozone H26n) is overlain by the thickest (~200 m) reverse-polarity interval (magnetozone H26r) thus far discovered in the Portland Formation, found in a total of six sites from the Stony Brook and Agawam sections. The highest analyzed part of the Portland Formation ends in normal polarity of magnetozone H27n, delineated by two sites in the Stony Brook section about 1800 m above the Hampden Basalt.

7. Chronostratigraphic Control

[25] Chronostratigraphic control for the Hartford basin nonmarine stratigraphic succession has traditionally been based on palynology, vertebrate biostratigraphy, and some inconclusive geochronology. The pre-CAMP New Haven

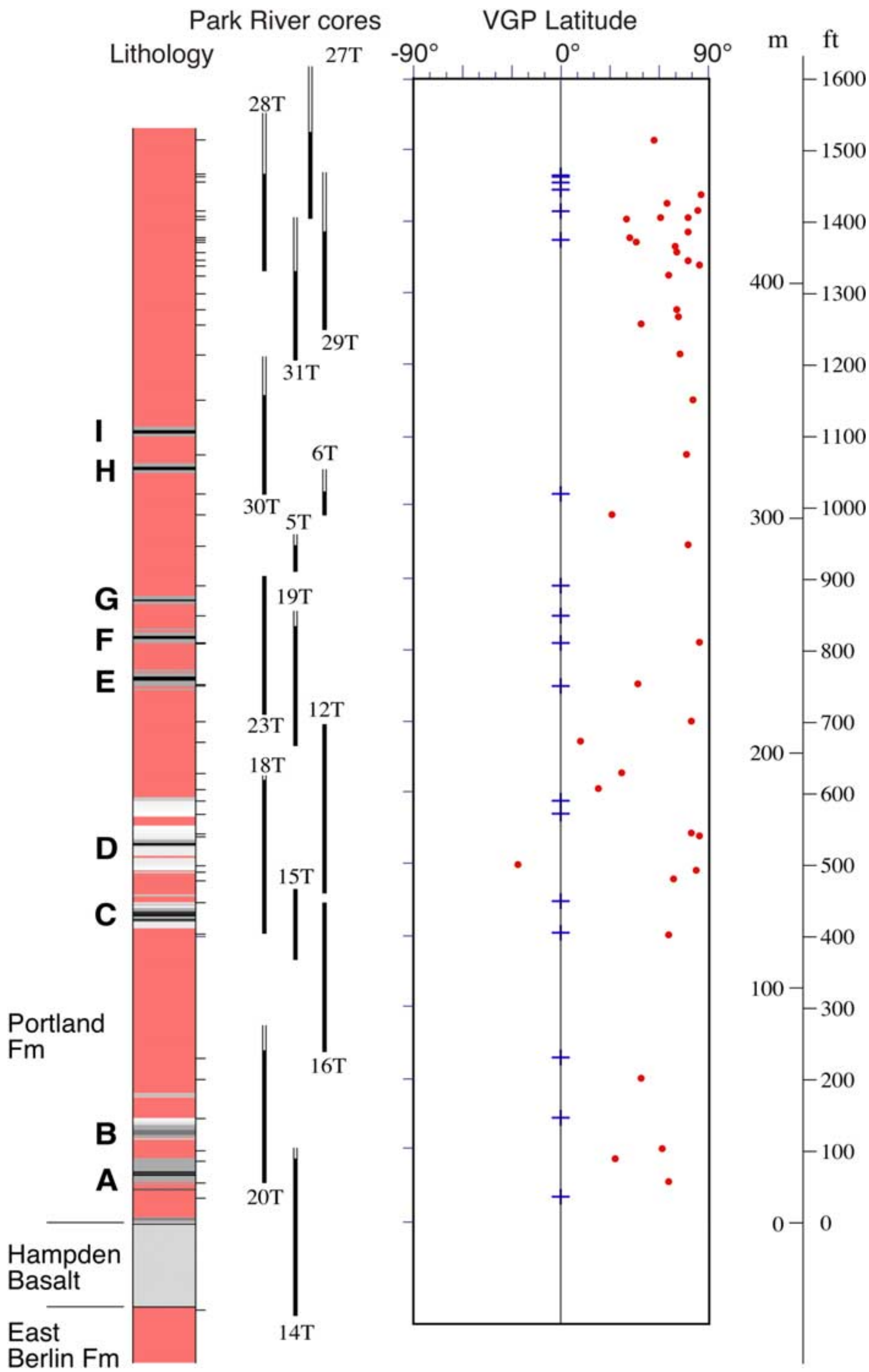

Figure 9. Composite section of the mostly covered interval of Smiths Ferry and Park River members of lower Portland Formation based on Park River cores taken by Army Corps of Engineers. Projection of bedding dips and tracing of distinctive beds labeled A to I were used to arrange the 14 cores in stratigraphic sequence. VGP latitudes for characteristic magnetizations of indirectly oriented samples were used to interpret polarity (values approaching +90° indicate normal polarity and values approaching −90° indicate reverse polarity; crosses along 0° axis indicate samples that did not provide interpretable data). See Figure 8 for integration with data from outcrop sections.

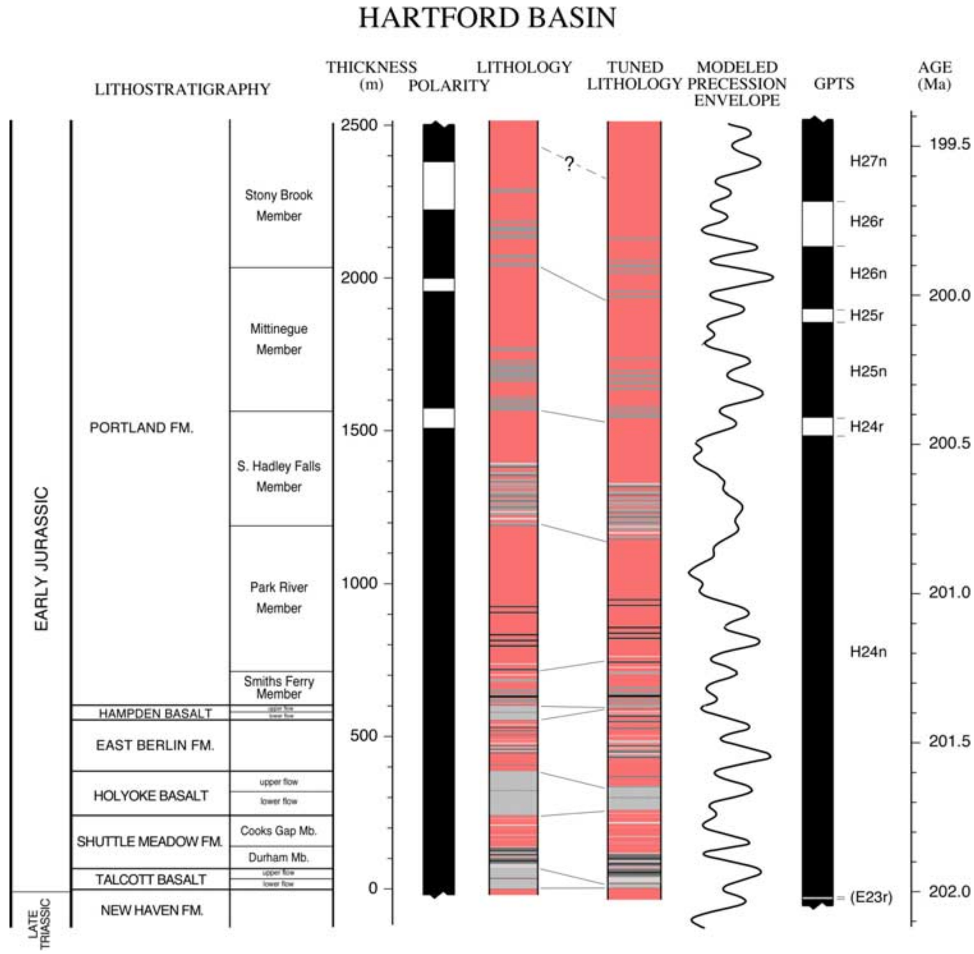

Figure 10. Astronomically tuned geomagnetic polarity timescale (GPTS) for the Early Jurassic based on cycle and magnetic polarity stratigraphy of composite section from Hartford basin. Conversion of stratigraphic thickness to age is based on interpolation within McLaughlin cycles, which are assumed to represent the 404-ka orbital eccentricity variation (modeled precession envelope described by *Whiteside et al.* [2007]), and indexed to an estimated age of 202 Ma for the Triassic-Jurassic boundary event. See Table 2.

Formation is almost entirely fluvial; a Late Triassic (latest Carnian to early Norian) palynoflorule was reported from the lower part of the unit [*Cornet*, 1977], which also contains reptile fossils of Late Triassic age [*Lucas et al.*, 1998; *Olsen et al.*, 2000]. A major advance was the recognition that a substantial portion of the Hartford section extended into the Jurassic. Palynoflorules in the cyclical lacustrine strata of the Shuttle Meadow Formation and the lower part of the Portland Formation are described as having Liassic (Early Jurassic) affinities [*Cornet et al.*, 1973; *Cornet and Traverse*, 1975]. The uppermost few meters of the New Haven Formation contains a palynoflorule of typical Early Jurassic aspect [*Heilman*, 1987; *Olsen et al.*, 2002b], indicating that the Triassic-Jurassic boundary must lie just below the base of the Talcott Basalt, in a homologous position with respect to the CAMP lavas in the Newark basin [*Olsen et al.*, 2002a].

[26] To date, the Hartford basin has yielded radioisotopic (K-Ar, $^{40}Ar/^{39}Ar$) dates that have widely scattered values (150–250+ Ma) attributed to variable argon loss and excess argon [*Armstrong and Besancon*, 1970; *Seidemann et al.*, 1984; *Seidemann*, 1988, 1989]. However, paleomagnetic and geochemical data suggest a one-to-one correspondence of the volcanic units in the Hartford basin to those in the Newark basin [*Prevot and McWilliams*, 1989] where geochronological efforts have been much more successful. The Palisade sill, a traditional target for radioisotopic dating, was probably a feeder for the lower extrusive unit (Orange Mountain Basalt) in the Newark basin [*Prevot and McWilliams*, 1989]. A U-Pb baddeleyite date of 201 ± 1 Ma [*Dunning and Hodych*, 1990] and a $^{40}Ar/^{39}Ar$ biotite date of 202.2 ± 1.3 Ma from a recrystallized sedimentary xenolith [*Sutter*, 1988] associated with the Palisade sill are consistent with an $^{40}Ar/^{39}Ar$ age of 201.0 ± 1.2 Ma on the Orange Mountain Basalt [*Hames et al.*, 2000]. Together these dates are compatible with an age somewhere in the range ∼200–202 Ma for the Triassic-Jurassic boundary that lies a few meters below the equivalent of the Orange Mountain Basalt in the Jacksonwald area of the Newark basin [*Olsen et al.*, 2002a]. In the Fundy basin of Nova Scotia, the North Mountain Basalt is a few meters above the Triassic-Jurassic boundary level [*Fowell and Traverse*, 1995; *Olsen et al.*, 2003] and has yielded U-Pb zircon dates of 202 ± 1 Ma [*Hodych and Dunning*, 1992] and 201.27 ± 0.27 [*Schoene et*

al., 2006] that strongly support the geochronology from the Newark basin section. More recently, zircon-bearing tuffs in association with ammonite-bearing strata in Peru yielded an U-Pb age of 201.58 ± 0.28 Ma for the marine Triassic-Jurassic boundary [*Schaltegger et al.*, 2008]. For consistency with our earlier work, we use the rounded-off integer value of 202 Ma for the Triassic-Jurassic boundary that has served as an anchor point in the astronomically tuned GPTS for the Late Triassic based on coring from the Newark basin [*Kent and Olsen*, 1999].

[27] The cyclical lacustrine strata of the Shuttle Meadow, East Berlin and lower Portland formations provide an opportunity for establishing an astronomical timescale for the Jurassic portion of the Hartford succession (from *Olsen et al.* [2005] and additional mapping for this paper). The Hartford extrusive zone encompasses about 600 ka, partitioned about equally between the Shuttle Meadow and East Berlin formations and assuming that there is comparatively little time represented in the basalt units [*Whiteside et al.*, 2007] (Figure 10). This duration is virtually identical to the cycle stratigraphic estimate for the duration of the Newark extrusive zone (580 ka [*Olsen et al.*, 1996b]), which is also within the resolution of available $^{40}Ar/^{39}Ar$ dating of the lowest and highest lava flows in the Newark basin [*Hames et al.*, 2000]. The pervasive and distinctive cyclic lithologic variation that reflects shallower and deeper lakes that we argue are tied to orbitally controlled precipitation/evaporation regimes continues into the lower Portland Formation. Key features are facies changes at the 20 ka precession cycle that are modulated by the short (100 ka) and long (400 ka) eccentricity cycles. Most usefully, the long eccentricity (sometimes referred to as the McLaughlin) cycle often corresponds to mapped lithologic members (e.g., Park River, South Hadley Falls, and Mittinegue members) and provides a basis for long-range correlation and a chronological framework. This is similar to the pattern in the Newark basin section [*Olsen*, 1986; *Olsen et al.*, 1996a], where a Triassic numerical timescale based on the fundamental Milankovitch periodicities [*Olsen and Kent*, 1996; *Kent and Olsen*, 1999] has been largely confirmed by high-precision geochronology [*Wang et al.*, 1998; *Furin et al.*, 2006].

[28] An astronomical timescale for the lower Portland Formation composite section indicates an average sediment accumulation rate of ∼1000 m/Ma (Figure 10). For comparison, the cyclical Late Triassic part of the Newark basin section accumulated at an average rate of only 160 m/Ma [*Olsen et al.*, 1996a; *Kent and Olsen*, 1999] although the Jurassic sedimentary units that are interbedded with the lavas can also have very high accumulation rates, for example, 26 m for the 20 ka cycle, corresponding to 1300 m/Ma, for the Towaco Formation [*Olsen et al.*, 1996b]. Lithologic expressions of 20 ka precession cyclicity are about 20 m thick in the Portland Formation (somewhat thinner in the Shuttle Meadow and East Berlin formations); the nominal average paleomagnetic site sampling interval of roughly 30 m thus represents a temporal resolution of one or two precession cycles. Even though the composite section for the Hartford basin has considerable uncertainties in the depth scale because it was assembled from various parts of the basin, only modest tuning (i.e., departures from sediment thickness as a first-order linear proxy of time) was required to account for the cyclicity by Milankovitch climate forcing. In all, the ∼2500 m of section from the base of the Talcott Basalt (base of the CAMP extrusive zone) into the lower to middle part of the Portland Formation represents ∼6 McLaughlin cycles, or ∼2.4 Ma of Early Jurassic time (Table 2).

Table 2. Astronomically Tuned Geomagnetic Polarity Timescale for ∼5 Ma Interval Centered on the Triassic-Jurassic Boundary Set at 202 Ma Based on Magnetic and Cycle Stratigraphy Data From the Hartford and Newark Basins[a]

Base of Unit	Depth, m	McLaughlin Cycle	Age, Ma
H27n	2386.95	66.814	199.687
H26r	2232.50	66.443	199.837
Stony Brook Member	2047.70	66	200.016
H26n	2008.60	65.918	200.049
H25r	1957.55	65.812	200.092
H25n	1578.10	65.019	200.412
Mittinegue Member	1568.79	65	200.420
H24r	1520.00	64.870	200.473
South Hadley Falls Member	1194.39	64	200.824
Park River Member	713.12	63	201.228
Smiths Ferry Member	379.64	62	201.632
Talcott (=Orange Mountain) Basalt	0	61.141	201.979
Tr/J (Newark basin)	−5.20	61.089	202
E24n = H24n	−10.70	61.036	202.021
VV (Exeter Member)	−12.3	61	202.036
E23r	−13.81	60.970	202.048
UU (Pine Ridge Member)	−59.50	60	202.440
TT	−124.7	59	202.844
E23n	−152.40	58.550	203.026
SS	−186.50	58	203.248
E22r	−197.66	57.796	203.330
E22n.2n	−230.95	57.215	203.565
E22n.1r	−232.56	57.188	203.576
RR	−241.30	57	203.652
E22n.1n	−288.37	56.284	203.941
QQ	−307.00	56	204.056
E21r.3r	−333.30	55.506	204.256
E21r.2n	−336.13	55.452	204.277
E21r.2r	−353.60	55.123	204.410
E21r.1n	−359.91	55.004	204.458
PP	−360.10	55	204.460

[a]Magnetic polarity chrons defined in Newark Basin Coring Project cores have the prefix E [*Kent and Olsen*, 1999] and those from Hartford basin outcrop composite section have prefix H (this study); polarity is designated by suffix n for normal and r for reverse. The base of each chron is given as the fractional position from the base of the enclosing McLaughlin member cycle, counted up from RaR-8 (informal cycle 1) in Stockton Formation to the Exeter Member (informal cycle 61) in the uppermost Passaic Formation in the Newark basin and continuing to Stony Brook Member (informal cycle 66) of Portland Formation in the Hartford basin. Depth in the Newark basin (negative numbers) is composite stratigraphic thickness scaled downward from base of Orange Mountain Basalt and normalized to Rutgers drill core based on successive core overlap correlations [*Olsen et al.*, 1996a]; depth in the Hartford basin (positive numbers) is measured upward from base of Talcott Basalt [this study; *Olsen et al.*, 2005]. Ages for the polarity chrons are based on interpolation within McLaughlin cycles, which are assumed to represent the 404-ka orbital eccentricity modulation of climate precession. The relative chronology is indexed to an estimated age of 202 Ma for the Triassic-Jurassic (Tr/J) boundary event. The depths of unit boundaries, interpolated values of position within a cycle, and ages are quoted with a precision needed for internal consistency.

8. Geomagnetic Polarity Timescale

[29] The Newark astronomically tuned GPTS was anchored to an age of 202 Ma for a level corresponding to the end-Triassic extinction level identified on the basis of

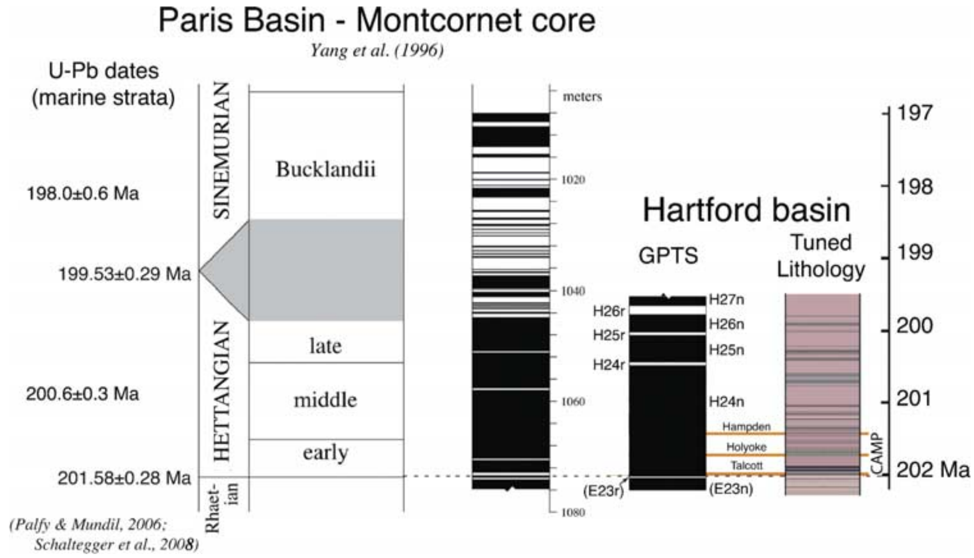

Figure 11. Magnetobiostratigraphy of Hettangian and Sinemurian marine sediments from the lower part of Montcornet core from the Paris basin plotted on a linear depth scale [*Yang et al.*, 1996] and compared to the astronomically tuned geomagnetic polarity timescale (GPTS) and lithology column for the Early Jurassic from the Hartford basin that was scaled in time using cycle stratigraphy and a Triassic-Jurassic boundary age of 202 Ma. Alignments of the Rhaetian-Hettangian (=Triassic-Jurassic) boundary level and the prominent reverse polarity chron H26r with an interval of predominantly reverse polarity between ~1041 and 1045 m in the Montcornet core would suggest that the Hettangian is only a few million years long. A short duration for the Hettangian is supported by U-Pb dates from early Sinemurian and Hettangian marine sediments with biostratigraphic control [*Pálfy and Mundil*, 2006; *Schaltegger et al.*, 2008].

palynology and supported by tetrapod footprint evidence [*Olsen et al.*, 1996a; *Kent and Olsen*, 1999; *Olsen et al.*, 2002b, 2003] that occurs about one precession cycle (~20 ka) before the Orange Mountain Basalt, the first CAMP lava in the Newark basin, and just after the end of Chron E23r, the last geomagnetic polarity reversal in the Triassic [*Kent and Olsen*, 1999] (Figure 10). Normal-polarity Chron E24n, which begins at 202.021 Ma, encompasses the igneous extrusive zone in the Newark basin and is correlative to the normal-polarity interval (Chron H24n) that encompasses the CAMP extrusive zone in the Hartford basin and extends to the uppermost part of the South Hadley Falls Member. This would make Chron H24n equal to nearly four McLaughlin cycles, or 1590 ka in total duration.

[30] Chron H24r is the first reverse-polarity interval in the Jurassic as recorded in the Hartford section; it is a thin (58 ± 50 m) magnetozone that occurs in the uppermost South Hadley Falls Member at about 200.45 Ma, or 1550 ka after the Triassic-Jurassic boundary according to the McLaughlin cyclicity. The overlying Mittinegue Member is mostly normal polarity (Chron H25n) until another thin (51 ± 27 m) reverse polarity magnetozone (Chron H25r) occurs in its youngest part at about 200.02 Ma. The succeeding Stony Brook Member has normal polarity (Chron H26n) in its lower part and a relatively thick (154 ± 15 m) reverse-polarity interval (Chron H26r) in its upper part. The sampled section ends in normal polarity of Chron H27n in what may still be the Stony Brook Member where the cyclic facies character fades; the transition from Chron H26r to Chron H27n occurs at an estimated age of 199.6 Ma and is the youngest polarity reversal delineated thus far in the Portland Formation.

[31] A plausible correlation can be made between the Hartford continental record and the magnetobiostratigraphy of Hettangian and Sinemurian sediments in the Montcornet core from the Paris basin [*Yang et al.*, 1996], the most detailed available marine record with magnetostratigraphy for the Early Jurassic (Figure 11). The ~30 m section (~1045–1075 m) of the Montcornet core that corresponds to the Hettangian according to biostratigraphy is characterized by predominantly normal polarity punctuated by several thin reverse polarity magnetozones, whereas reverse polarity is more prevalent above about 1045 m in the late Hettangian to early Sinemurian. The Montcornet polarity pattern thus suggests a correlation of Chrons H24n to H26n to the predominantly normal polarities of the Hettangian, and Chron H26r, the longest and youngest reverse polarity interval thus far delineated in the Early Jurassic of the Hartford basin section, to the mostly reverse polarity interval (~1041–1045 m) in sediments designated as late Hettangian to early Sinemurian in the Montcornet core. This general correlation implies that the nearly 600 m thick CAMP extrusive zone of interbedded continental sediments and lavas and the nearly 2000 m thick overlying section of continental sediments of the lower Portland Formation all accumulated during the Hettangian. The ~2.4 Ma duration estimated from cycle stratigraphy for this part of the Hartford basin section is consistent with its correlation to the Hettangian, which recent U-Pb dating indicates is only 2 to 3 Ma long [*Pálfy and Mundil*, 2006; *Schaltegger et al.*, 2008].

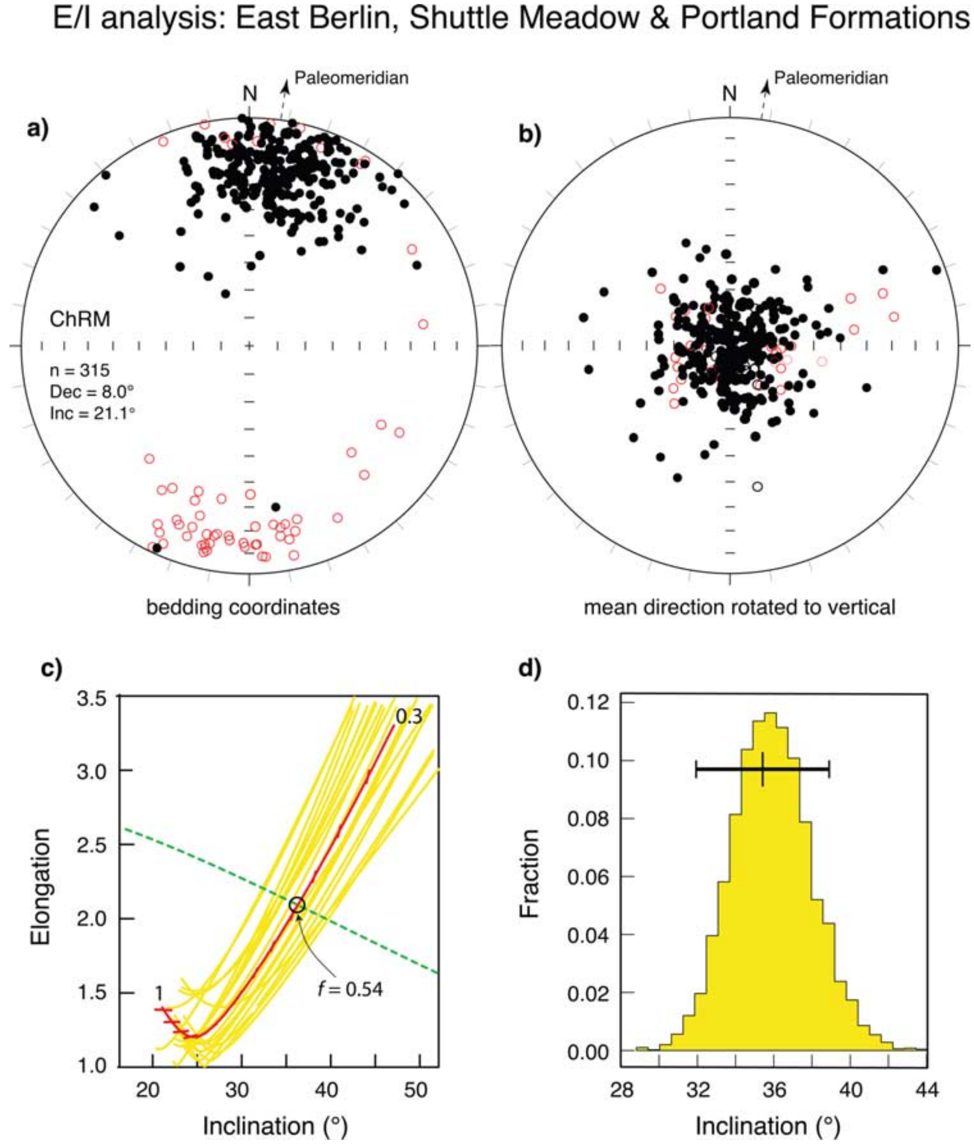

Figure 12. Equal-area projections of (a) sample ChRM directions after bedding tilt correction from East Berlin, Shuttle Meadow, and Portland formations and (b) the same distribution rotated so that the overall mean direction (Table 1) corresponds to the vertical axis to view better the shape of the distribution, which is elongated perpendicular to the paleomeridian and indicative of inclination flattening. (c) E/I analysis [*Tauxe and Kent*, 2004] of the sample ChRM directions from East Berlin, Shuttle Meadow, and lower Portland formations with the trajectory of mean inclination versus elongation of the distribution calculated as the data are inverted with values for the flattening factor (f) ranging from 0.3 to 1.0. The predicted E/I trend of the TK03.GAD geomagnetic field model is shown as dashed line; the E/I of the data consistent with the model is circled and corresponds to $f = 0.54$. (d) Histogram of 1000 intersections of the kind shown in Figure 12c from bootstrapped curves. The mean and 95% confidence bounds of the corrected inclination are shown (see Table 3).

[32] The ~30 m Hettangian interval of the Montcornet core is also interpreted to have five very thin reverse polarity intervals [*Yang et al.*, 1996]. One of the thin reverse intervals in the middle Hettangian and another in the late Hettangian might correspond to the short subchrons H24r and H25r, respectively, in the Hartford section (Figure 11). However, no convincing counterparts have yet been found in the Hartford (or Newark) basin section for any of the three thin reverse polarity intervals in the early Hettangian part of the Montcornet core, which ostensibly should correspond in age to the CAMP extrusive zone. Some of the thin polarity intervals in the Montcornet core might be artifacts of inverted or overprinted core segments; alternatively, short polarity intervals remain to be discovered in the CAMP extrusive zone of the Newark and Hartford basins.

9. Inclination Shallowing

[33] The mean inclination for the sedimentary units (22.2 ± 3.7° for 71 sites, or 21.1 ± 2.1° for 315 samples from the

Table 3. Results of E/I Analysis on ChRM Sample Directions From the Early Jurassic Shuttle Meadow, East Berlin and Lower Portland Formations From the Hartford Basin[a]

Locality	Slat, °N	Slon, °E	Age, Ma	n	Dec, deg	Inc, deg	λ, °N	f	Inc′, deg	±Inc′, deg	λ', °N	$\pm\lambda'$, deg
Hartford basin	42.0	−72.5	201	315	8.0	21.1	10.9	0.54	35.5	32–39	19.2	17.4–22.0

[a]Slat and Slon are the latitude and longitude of the nominal sampling locality; Age is the nominal mean age of the early Jurassic rock units; n is the number of data analyzed; Dec and Inc are the mean declination and mean inclination of the characteristic magnetization data; λ is paleolatitude calculated with dipole formula from the mean inclination; f is the flattening factor determined from E/I analysis; Inc′ is the corrected mean inclination and ±Inc′ is its 95% confidence interval; λ' is the corresponding corrected paleolatitude and $\pm\lambda'$ is its 95% confidence interval.

Shuttle Meadow, East Berlin, and Portland formations) is significantly shallower than the mean inclination of 33.9 ± 8° for the closely associated CAMP volcanic units (calculated from the mean paleopole of 66.3°N 97.3°E A_{95} = 5° for the three extrusive units in the Newark and Hartford basins [*Prevot and McWilliams*, 1989]). This discrepancy may simply be an artifact of unaveraged paleosecular variation in the CAMP volcanics or more likely is an indication of sedimentary inclination error, which can occur during deposition of hematite-bearing sediments [*Tauxe and Kent*, 1984]. A similar discrepancy was observed in the Newark basin where the elongation/inclination (E/I) technique [*Tauxe and Kent*, 2004] was successfully used to detect and correct inclination error in the characteristic directions from the sedimentary units [*Kent and Tauxe*, 2005].

[34] The E/I technique was applied to a data set of 315 sample ChRM directions from the Shuttle Meadow, East Berlin and Portland formations, which represent 2.4 Ma of the earliest Jurassic. This data set should be sufficiently large in number of samples and length of record to capture the full range of secular variation yet still short enough to avoid introducing a bias in the directional distribution from polar wander. The distribution of ChRM directions is found to be elongated east-west, perpendicular to the paleomeridian, that is, flattened in the paleohorizontal or bedding plane (Figures 12a and 12b). In contrast, statistical geomagnetic field models using a giant Gaussian process (e.g., CP88.GAD [*Constable and Parker*, 1988] and TK03.GAD [*Tauxe and Kent*, 2004]) predict not only that the mean field inclination, I, is a function of latitude, λ:

$$\tan \mathrm{I} = 2 \tan \lambda \qquad (1)$$

but also that secular variation results in a distribution of virtual geomagnetic poles (longitudes and latitudes) that is

Table 4. Paleomagnetic Poles From Selected Late Triassic and Earliest Jurassic Rocks From North America[a]

Locality	Age, Ma	f	Plat′, °N	Plon′, °E	A_{95}, deg	Ref
Sedimentary Results Corrected for Inclination Error						
Hartford basin						
SM + EB + PF (H)	201	0.54	66.6	88.2	2.3	1
Newark Basin Coring Project						
Martinsville (M)	204	0.66	67.8	96.1	2.9	2, 3
Weston Canal (W)	207	0.49	66.6	86.5	2.9	2, 3
Somerset (S)	211	0.49	61.7	95.3	2.0	2, 3
Rutgers (R)	214	0.66	60.1	97.1	1.4	2, 3
Titusville (T)	217	0.63	59.9	99.5	1.7	2, 3
Nursery Road (N)	221	0.40	60.5	101.6	2.5	2, 3
Princeton (P)	227	0.56	54.2	106.6	2.0	2, 3
Dan River basin (D)	221	0.59	58.5	99.8	1.1	3, 4
Colorado Plateau						
Moenave+Wingate (mo+wi)	201	1.0[b]	59.3	59.0	8.3	5, 6
Moenave+Wingate	201	0.5[c]	65.3	57.0	8.3	1, 5, 6
Plus 13.5° counterclockwise rotation (Mo+Wi)	201	0.5[c]	66.3	85.9	8.3	1, 5–7
Igneous Results						
Newark+Hartford lavas (CAMP)	201		66.3	97.3	5.0	8, 9
Manicouagan (MI)	214		58.8	89.9	5.8	10–12

[a]Age is the nominal age of the Late Triassic and Early Jurassic rock units; f is the flattening factor determined from E/I analysis; Plat′ and Plon′ are the latitude and longitude, and A_{95} is the radius of the 95% circle of confidence, of the paleopole that corresponds to corrected directions for the sedimentary results (see also Table 3). Ref is the literature source for the age and paleomagnetic data: 1, This study; 2, *Kent et al.* [1995]; 3, *Kent and Tauxe* [2005]; 4, *Kent and Olsen* [1997]; 5, *Ekstrand and Butler* [1989]; 6, *Molina-Garza et al.* [2003]; 7, *Kent and Witte* [1993]; 8, *Prevot and McWilliams* [1989]; 9, *Hames et al.* [2000]; 10, *Robertson* [1967]; 11, *Larochelle and Currie* [1967]; and 12, *Hodych and Dunning* [1992]. The 201 Ma reference pole position for stable North America (Figure 13) is the mean of corrected Hartford basin sediments (H), corrected Martinsville core sediments (M), and Newark+Hartford lavas (CAMP) at 67.0°N 93.8°E (A_{95} = 3.2°).

[b]Uncorrected data.

[c]Corrected data with an assumed flattening factor for Moenave+Wingate.

essentially circular at any observation site, implying that the distribution of directions (declinations and inclinations) will be systematically more elongate north-south, along the paleomeridian, as the observation site latitude decreases from the pole(s) to the equator. If the directions were affected by inclination error (either during deposition or imparted by compaction), the observed inclination, I_o, will be related to the ambient field inclination, I_f, by:

$$\tan I_o = f \tan I_f \tag{2}$$

where f is the flattening factor [e.g., *King*, 1955]. Inclination error affects the distribution of directions by increasing the east-west elongation while decreasing inclination. If inclination error is the cause of the shallow bias, the data can be inverted using the inverse of equation (2) searching for a value of f that yields an E/I combination that is consistent with the field model; the corrected mean inclination should provide a more accurate estimate of latitude according to equation (1). The hypothesis that the statistical properties of the geomagnetic field in remote epochs were similar to the more recent (0–5 Ma) geomagnetic field was supported by the ability of the E/I method to produce an internally consistent latitudinal framework in the Late Triassic from studies made over a broad region [*Kent and Tauxe*, 2005].

[35] E/I analysis of the Hartford sedimentary ChRM data produces a result consistent with the geomagnetic field model at a mean flattening factor of 0.54, which is well within the range of f values determined by E/I analysis of other data sets of similar hematite-bearing sedimentary rocks [*Kent and Tauxe*, 2005]; the corrected inclination is 35.5°, with bootstrapped 95% confidence limits of 32° and 39° (Figures 12c and 12d and Table 3). The corrected inclination is more than 14° steeper than the uncorrected mean inclination, in keeping with what *Tan et al.* [2007] found with their magnetic anisotropy correction for the Passaic Formation red beds in the Newark basin. The corrected Hartford direction is also not significantly different from the mean inclination of 33.9 ± 8° estimated from the mean pole for the Newark and Hartford CAMP volcanic units [*Prevot and McWilliams*, 1989], although the volcanics pole represents only a small number of cooling units and thus may not adequately average paleosecular variation.

10. Paleomagnetic Poles and J1 Cusp

[36] The (north) paleopole corresponding to the corrected Hartford sedimentary ChRM direction is located at 66.6°N 88.2°E A_{95} = 2.3°; this is only an insignificant 3.6° away from the Newark and Hartford CAMP volcanics pole at 66.3°N 97.3°E A_{95} = 5° [*Prevot and McWilliams*, 1989] and only an insignificant 3.3° from the paleopole for corrected results from latest Triassic and earliest Jurassic sediments in the Martinsville core immediately below the CAMP extrusive zone in the Newark basin (67.8°N 96.1°E A_{95} = 2.9° [*Kent and Tauxe*, 2005]) (Figure 13). A comparison of paleopoles of tightly correlated Newark Supergroup strata of Late Triassic age from the widely separated Newark basin and the Dan River basin of North Carolina and Virginia shows no evidence of vertical axis tectonic rotations [*Kent and Olsen*, 1997]. Similarly, the positive fold test demonstrated here for data from Early Jurassic strata in the Newark and Hartford basins is further indication that these early Mesozoic rift basins have maintained tectonic coherence with respect to eastern North America. Accordingly, we regard the average of the paleopoles for the CAMP volcanics, Newark corrected sediments and Hartford corrected sediments (67.0°N 93.8°E A_{95} = 3.2°, N = 3) as representative of the ~201 Ma pole position for at least eastern North America.

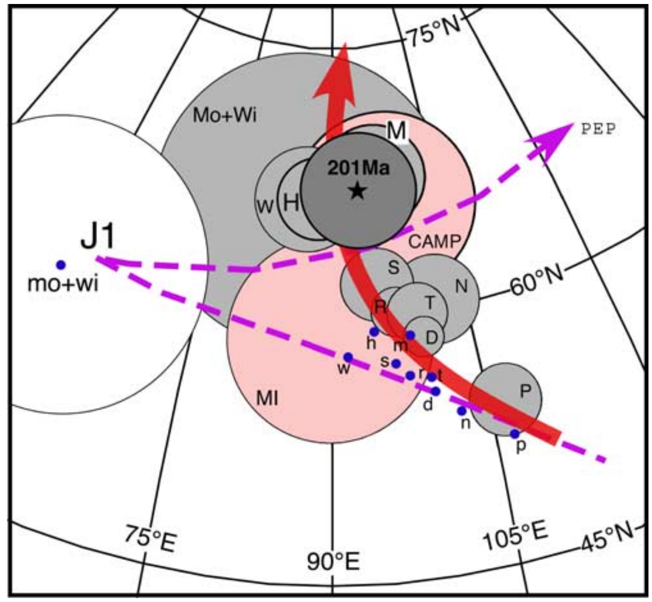

Figure 13. Selected Late Triassic and Early Jurassic paleopoles from North America. Large circles are A_{95} for igneous poles (CAMP lavas; MI, Manicouagan impact structure) and poles corrected for inclination error: P, Princeton; N, Nursery; T, Titusville; R, Rutgers; S, Somerset; W, Weston Canal; and M, Martinsville for NBCP cores; D for Dan River basin, H for Hartford basin (see Table 4), whereas small filled circles with lowercase letters are paleopoles for same sedimentary units before correction for inclination error. Mean paleopole for Moenave and Wingate formations is shown with circle of confidence labeled mo+wi, and as shaded circle of confidence labeled Mo+Wi after correction for inclination error (assumed flattening factor, f = 0.5) and 13.5° net clockwise rotation of Colorado Plateau with respect to eastern North America (Table 4). Star with circle of confidence labeled "201 Ma" is mean paleopole of CAMP lavas and corrected Martinsville core (M) and Hartford basin (H) sedimentary directions. J1 is calculated position of the ~203 Ma cusp joining Permian-Triassic and Jurassic-Cretaceous tracks from a paleomagnetic Euler pole model (path shown by line with arrow labeled PEP [*Gordon et al.*, 1984]) that now is seen to be biased by inclination error and rotation of the Colorado Plateau. Heavier line with arrow of time is our preferred empirical APW path for North America for the Late Triassic and Early Jurassic based on igneous results and sedimentary data corrected for inclination error from eastern North America.

[38] The paleopole from the Moenave Formation on the Colorado Plateau is usually considered to practically define the J1 cusp [e.g., *Molina-Garza et al.*, 1995]. A mean pole position based on modern studies of the Moenave Formation at Vermillion Cliffs in Arizona and Utah [*Ekstrand and Butler*, 1989] and at Echo Cliffs in Arizona [*Molina-Garza et al.*, 2003], and of its presumed lateral equivalent, the Wingate Formation at Comb Ridge in Utah [*Molina-Garza et al.*, 2003], is located at 59.3°N 59.0°E A_{95} = 8.3° (N = 3), which is indeed very close to the canonical location of the J1 cusp derived from the paleomagnetic Euler pole model of APW for North America (60.5°N 62.4°E [*Gordon et al.*, 1984]) (Figure 13). Available biostratigraphic evidence from palynoflora and tetrapod fossils and footprints suggest that the Moenave and Wingate formations of the Glen Canyon Group are latest Triassic (Rhaetian) to earliest Jurassic (Hettangian) in age [*Molina-Garza et al.*, 2003]. This is virtually the same time interval encompassed by latest Triassic sediments from the Martinsville core in the Newark basin, earliest Jurassic sediments from the Hartford basin, and the CAMP extrusives in the Newark and Hartford basins. However, the Moenave/Wingate mean pole differs by a significant 17.2° of great circle arc from the coeval 201 Ma mean pole from the Newark/Hartford basins.

[39] When projected to a nominal sampling location for the Moenave/Wingate on the Colorado Plateau (36.5°N 111°W), the 201 Ma Newark/Hartford pole predicts a paleomagnetic direction (D = 350.2° I = 28.5°) that is considerably more northwesterly but also much steeper than the Moenave/Wingate direction (D = 5.1° I = 12.2°). The 14.9° difference in declination can be attributed to net clockwise rotation of the Colorado Plateau relative to eastern North America by a comparably large net amount (see summary of plateau rotation estimates by *Steiner and Lucas* [2000]). Clockwise plateau rotation, however, cannot also account for the 16.3° shallower direction in the Moenave/Wingate, which is more likely ascribed to inclination error. Flattening factors in the range 0.4 to 0.6 have been found for characteristic magnetizations in other red bed units [e.g., *Kent and Tauxe*, 2005] and if applicable to the Moenave/Wingate magnetizations, a correction using a nominal value of $f \sim 0.5$ would steepen the Moenave/Wingate inclination to 23.4° and bring it into good agreement (within the error limits) with the predicted inclination from the Newark/Hartford 201 Ma pole whereas a correction using $f = 0.4$ would steepen the Moenave/Wingate inclination to 28.4° and make it agree almost precisely with the predicted inclination from the coeval Newark/Hartford pole.

[40] The inferred degree of inclination error in the Moenave/Wingate magnetizations obviously needs to be verified by E/I analysis [*Tauxe and Kent*, 2004] or by the anisotropy method [*Tan et al.*, 2007]. At this juncture, we find it intriguing that a correction of the Moenave/Wingate mean direction for 13.5° clockwise rotation of the Colorado Plateau [*Kent and Witte*, 1993] and for inclination error corresponding to a nominal flattening factor of $f \sim 0.5$ yields a paleopole at 66°N 86°E that is in excellent agreement with the coeval Newark/Hartford 201 Ma paleopole. The apparent concordance between the corrected mean Moenave/Wingate and coeval Newark/Hartford 201 Ma poles and their appreciable departure from the postulated position of the J1 cusp suggest that this key feature of paleomagnetic Euler pole analysis is largely an artifact of Colorado Plateau rotation and sedimentary inclination error. Instead, the corrected data show that APW for North America proceeds in a more northerly direction to higher latitudes over the Late Triassic and Early Jurassic (Figure 13).

11. Discussion

[41] Chron E24n (= H24n) represents the thickest polarity unit in the Newark Supergroup polarity sequence, encompassing about 1600 m of section in the Hartford basin and a minimum of 1000 m in the Newark basin (Figure 8). Chron E24n began just ~40 ka prior to the earliest CAMP lavas in the Newark basin [*Kent and Olsen*, 1999], extended over the 600-ka-long CAMP extrusive zone in both the Newark and Hartford basins [*Olsen et al.*, 1996b; *Whiteside et al.*, 2007], and apparently ended with the first reverse polarity interval of the Jurassic (Chron H24r) that occurred 950 ka after the last CAMP lavas in the Hartford basin (Figure 10). Nevertheless, the 1590 ka duration of Chron H24n is not the longest in the Late Triassic–earliest Jurassic GPTS, being exceeded by three Triassic polarity chrons: 2003 ka for E11r, 1797 ka for E16n, and 1618 ka for E8r [*Kent and Olsen*, 1999]. Moreover, the first six polarity intervals of the early Jurassic (H24n to H26r), which range in duration from ~50 ka to 1590 ka and represent ~2400 ka according to cycle stratigraphy (Table 2), have an average duration of around 400 ka, which is shorter than the average duration of 530 ka for polarity intervals over the preceding 30 Ma of the Late Triassic [*Kent and Olsen*, 1999]. These data do not support the existence of a polarity superchron, or even a marked decrease in geomagnetic reversal frequency in the Early Jurassic. The fact that 80% of the last 2.5 Ma of the Late Triassic and first ~2.5 Ma of the Early Jurassic had normal geomagnetic polarity is noteworthy but its significance is unclear since there are several other 5 Ma intervals with a strong polarity bias in the Newark GPTS, for example, the interval between 205 and 210 Ma has 80% reverse polarity [*Kent and Olsen*, 1999].

[42] Although the geomagnetic polarity column in the recent geologic timescale [*Gradstein et al.*, 2004] depicts the earliest Jurassic as having predominantly reverse polarity based on preliminary data from Austria [*Steiner and Ogg*, 1988], latest Triassic and earliest Jurassic time is in fact characterized by predominantly normal geomagnetic polarity based on published magnetostratigraphies from marine sections (St. Audrie′s Bay [*Hounslow et al.*, 2004] and the Montcornet core from the Paris basin [*Yang et al.*, 1996]) that are consistent with the Hartford data. The St. Audrie's Bay and especially the Montcornet magnetostratigraphic records are punctuated by a number of relatively thin magnetic zones that have been interpreted as representing short reverse polarity intervals [*Yang et al.*, 1996; *Hounslow et al.*, 2004]. The shortest reverse polarity intervals in the Newark and Hartford polarity sequence are Chrons E23r (~20 ka) and H24r (at least 10 ka but perhaps as long as 100 ka; Table 2), which happen to bracket the long normal polarity Chron E24n/H24n that includes CAMP volcanism. One or more levels with anomalous paleomagnetic directions in the Moroccan record of CAMP rocks have been correlated to Chron E23r, suggesting that CAMP volcanism started prior to the end of the Triassic

[*Knight et al.*, 2004; *Marzoli et al.*, 2004]. However, the anomalous directions from Morocco taken at face value could just as well reflect one of the short reverse intervals or polarity excursions in the Hettangian of the Montcornet core from the Paris basin [*Yang et al.*, 1996] that, if real, remain to be identified in the Newark and Hartford CAMP interval. Additional short polarity intervals may exist in the Hartford (and Newark) basin sections although given the present sampling density, any new polarity intervals would not be expected to be longer than about 20 ka. In the Cenozoic geomagnetic polarity record, such short features might qualify as reversal excursions or polarity fluctuations, rather than full polarity chrons, and be difficult to use for global correlation [e.g., *Krijgsman and Kent*, 2004; *Lowrie and Kent*, 2004]. The isolated sample with southerly VGP latitude about 100 m (or ~100 ka) above the Hampden Basalt (Figure 9) might be an example of such a reversal excursion that obviously needs to be verified.

[43] The principal exemplars of the J1 cusp, the paleopoles from the Moenave Formation and laterally correlative Wingate Formation from the Colorado Plateau, can be reconciled to the coeval 201 Ma reference paleopole for eastern North America (67.0°N 93.8°E A_{95} = 3.2°) based on data from the Newark/Hartford CAMP lavas and sedimentary units corrected for inclination error by net clockwise rotation of the Colorado Plateau and inclination error. The resulting APW path follows a more northerly trend that effectively bypasses the prominent J1 cusp of the paleomagnetic Euler pole model of North American APW [*Gordon et al.*, 1984; *May and Butler*, 1986]. The prevalence of inclination error indicated by comparisons to coeval igneous data [*Gilder et al.*, 2003] and by E/I analysis [*Krijgsman and Tauxe*, 2004; *Tauxe and Kent*, 2004; *Kent and Tauxe*, 2005] and the anisotropy technique [*Tan and Kodama*, 2002; *Tan et al.*, 2003, 2007] will require a comprehensive reevaluation of paleomagnetic data from sedimentary rocks used for paleopole and paleolatitudinal studies, including attempts at using paleomagnetic Euler pole analysis for estimating Colorado Plateau rotation [*Bryan and Gordon*, 1990]. In the meantime, we suggest that the APW path for the eastern (stable) part of North America may best be delineated by taking into consideration results from Jurassic igneous units from the White Mountain magma series in New England, including a venerable result from the ~169 Ma Belknap Mountains and Mount Monadnock plutons (85.4°N 354.6°E A_{95} = 3.5°; *Opdyke and Wensink* [1966], as recalculated by *Van Fossen and Kent* [1990] and confirmed by them with thermal demagnetization results giving a paleopole that includes the White Mountain batholith at 88.4°N 82.1°E A_{95} = 6.1°) and the dual-polarity paleopole for the 169 Ma Moat Volcanics (81.6°N 89.7°E A_{95} = 6° [*Van Fossen and Kent*, 1990]), which would advance an APW path for the Mesozoic that is reminiscent of the high-latitude route proposed by *Irving and Irving* [1982]. Parenthetically, virtually all of the other dozen or so listings of igneous results deemed reliable for the Jurassic of North America in a recent paleopole compilation [*Besse and Courtillot*, 2002] have a wide cited age range, from the 180 Ma dikes in the Piedmont of North Carolina [*Smith*, 1987] to the 201 Ma Newark Supergroup volcanics [*Prevot and McWilliams*, 1989], but more likely are poorly dated entries for the same short-lived (<1 Ma) CAMP event at around 201 Ma. Even more recent radioisotopic ages for CAMP igneous rocks are spread over a ~10-Ma interval [*Marzoli et al.*, 1999; *Knight et al.*, 2004]. Scatter in ages for CAMP (and indirectly, Chron E24n) may also help account for the unverified long interval of low reversal frequency and normal polarity in the Early Jurassic that appears in compilations of paleomagnetic polarity data [e.g., *Johnson et al.*, 1995; *Algeo*, 1996].

[44] Finally, the new earliest Jurassic (201 Ma) reference paleopole based on the Newark/Hartford data provides accurate and precise paleolatitudinal control, which is fully consistent with Late Triassic data corrected for inclination error from North America and other North Atlantic-bordering continents in a Pangea reconstruction [*Kent and Tauxe*, 2005]. By the Early Jurassic, much of North America had drifted northward into the arid belt, in good agreement with paleoclimate indicators of aridity such as eolian sandstones in the Pomperaug basin of Connecticut (projected paleolatitude ~19°N) in eastern North America [*LeTourneau and Huber*, 2006] and the appearance of prominent eolian sandstones in rock units of the Glen Canyon Group (American Southwest) from the Wingate Formation (projected paleolatitude ~15°N) that culminated in deposition of the Navajo Sandstone, one of the largest ergs on Earth [*Blakey et al.*, 1988]. Any global climatic effects of CAMP will need to be evaluated in the context of geographically distinct climate changes reflecting continental drift through latitudinal climate belts.

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