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Invited review

Astrochronostratigraphic polarity time scale (APTS) for the Late Triassic and Early Jurassic from continental sediments and correlation with standard marine stages

Dennis V. Kent [a,b,*], Paul E. Olsen [b], Giovanni Muttoni [c]

[a] Earth and Planetary Sciences, Rutgers University, Piscataway, NJ 08854, USA
[b] Lamont-Doherty Earth Observatory of Columbia University, Palisades, NY 10964, USA
[c] Dipartimento di Scienze della Terra 'Ardito Desio', Università degli Studi di Milano, via Mangiagalli 34, I-20133 Milan, Italy

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ABSTRACT

Paleomagnetic and cycle stratigraphic analyses of nearly 7000 m of section from continuous cores in the Newark basin and an overlapping 2500 meter-thick composite outcrop and core section in the nearby Hartford basin provide an astrochronostratigraphic polarity time-scale (APTS) for practically the entire Late Triassic (Carnian, Norian and Rhaetian) and the Hettangian and early Sinemurian stages of the Early Jurassic (233 to 199 Ma in toto). Aperiodic magnetic polarity reversals make a distinctive pattern of normal and reverse chrons for correlation, ideally paced by the periodic timing of orbital climate cycles, and anchored to million years ago (Ma) by high-precision U-Pb zircon dates from stratigraphically-constrained basalts of the Central Atlantic Magmatic Province (CAMP). Pinned by the CAMP dates, the Newark-Hartford APTS is calibrated by sixty-six McLaughlin cycles, each a reflection of climate forcing by the long astronomical eccentricity variation with the stable 405 kyr period, from 199.5 to 225.8 Ma and encompassing fifty-one magnetic polarity intervals, making it one of the longest continuous astrochronostratigraphic polarity time-scales available in the Mesozoic and Cenozoic. Extrapolation of sediment accumulation rates in fluvial sediments in the basal Newark section extends the sequence an additional fifteen polarity intervals to 232.7 Ma. The lengths of the 66 polarity chrons vary from 0.011 Myr (Chron E23r) to 1.63 Myr (Chron H24n) with an overall mean duration of 0.53 Myr. The oldest CAMP basalts provide a zircon U-Pb-based estimated age of 201.5 Ma for the base of the stratigraphically superjacent McLaughlin cycle 61 and 201.6 Ma using cycle stratigraphy for the onset of the immediately subjacent Chron E23r. The calibration age of 201.5 Ma for the base of McLaughlin cycle 61 is remarkably consistent with the calculated phase of the 498th long eccentricity cycle counting back using a period of 405 kyr from the most recent peak at 0.216 Ma. Accordingly, we suggest a nomenclature (Ecc405:k, where k is the cycle number or fraction thereof) to unambiguously assign ages from the astrochronostratigraphy. Magnetostratigraphic correlation of key Tethyan sections with diagnostic marine biostratigraphic elements to the Newark-Hartford APTS allows determination of numerical ages of standard marine stages, as follows: 227 Ma for the Carnian/Norian boundary, 205.5 Ma for the Norian/Rhaetian boundary (using a chemostratigraphic criterion, or about 4 Myr older for alternative criteria), 201.4 Ma for the Triassic/Jurassic boundary, and 199.5 Ma for the Hettangian/Sinemurian boundary. These age estimates are in excellent agreement with available constraints from high-precision U-Pb zircon dating from the Pucara Basin of Peru and along with the presence of the short Chron E23r in several basins argue strongly against suggestions that millions of years of Rhaetian time is missing in a cryptic hiatus or unconformity that supposedly occurs just above Chron E23r in the Newark Supergroup basins. It is more parsimonious to explain the apparent temporal delays in appearances and disappearances of palynoflora, conchostracans, and other endemic taxa in continental deposits as a reflection of demonstrated continental drift across climate belts and the misinterpretation of ecostratigraphy as chronostratigraphy. The Newark-Hartford APTS provides a chronostratigraphic template for continuing efforts at correlation of Late Triassic and Early Jurassic continental and marine sections throughout the world, including integration with atmospheric pCO_2 measurements from paleosol carbonates and carbon isotopic measurements from marine carbonates to better understand the global carbon cycle as well as understanding the causes of and recovery from the end-Triassic mass extinction.

* Corresponding author at: Earth and Planetary Sciences, Rutgers University, Piscataway, NJ 08854, USA.
E-mail address: dvk@rutgers.edu (D.V. Kent).

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Contents

1. Introduction

Recently compiled geological time-scales (GTS) have had particular difficulty with Triassic chronology. This is especially evident from the wide disparity in age estimates for the Carnian/Norian boundary, which changed from 216.5 Ma (Mega-annum or million years ago) in GTS2004 (Gradstein et al., 2004) to 228 Ma in GTS2012 (Gradstein et al., 2012). The difference of >10 Myr (million years in duration) made it one of the largest boundary age revisions in the Phanerozic in these time-scales. The change was largely due to avoidance in GTS2004 of U-Pb dates from the Latemar series of Ladinian age (which, ironically, had perhaps more U-Pb dated levels than any stage in the Triassic) that disagreed with the then-prevalent perception of a low sedimentation rate (50 m/Myr) and long duration (about 12 Myr) for the Latemar. This compared to better-substantiated values a full order of magnitude higher and shorter (500 m/Myr and only 1 Myr; see decisive numerical analysis by Meyers (2008)) and had the effect of squeezing or shortening other stages in the Triassic, like the Carnian. An updated chronology as summarized in Fig. 1 and described in detail below now makes the Late Triassic (Carnian, Norian, Rhaetian stages), at >35 Myr in duration, one of the longest epochs in the Phanerozoic.

Despite its considerable duration, there are few direct age calibrations for Late Triassic marine sections with biostratigraphy, which are traditionally used to define standard geological stages, and the available dates provide a geochronological resolution of only 5–10 Myr at best. The widely studied fossiliferous albeit condensed Hallstatt marine facies offer a better biochronological resolution of around 3 Myr (4 biozones with about 11 subzones) but the common presence of unconformities (mostly hardground hiatuses) limits the use of sediment thickness as a first-order proxy of time for interpolation and development of a reliable composite sequence (e.g., Krystyn et al., 2002). A continental time-scale based on tetrapod biochronology has even lower resolution of about 8 Myr with only two to four land vertebrate faunachrons recognized in the entire Late Triassic (Lucas, 2010). Direct Triassic marine-terrestrial links are even more limited in number, essentially confined to the mass extinction in both marine and nonmarine biota just prior to the Triassic-Jurassic boundary (TJB) (Benton et al., 2014). In continental rift basins of the now-dispersed Pangea supercontinent, the TJB has been inferred to be in close temporal proximity to the older lavas of the Central Atlantic Magmatic Province (CAMP), which also provide key high-precision U-Pb dates for time-scale calibration (Blackburn et al., 2013). The precise timing of global events leading up to the end-Triassic extinctions and biotic recovery into the Early Jurassic is of interest to constrain the possible cause(s) of the extinction event, its development in different fossil groups and environments, and the pace and nature of the biotic recovery. More broadly, an accurate depiction of the timing of processes reflected in the rock record is invaluable to gauge the intermingled roles of paleogeography and greenhouse gases on global climate (Goddéris et al., 2008; Schaller et al., 2015) and the distribution of biota (e.g., Kent et al., 2014; Whiteside et al., 2011a, 2011b).

Our strategy for developing a temporal template for global correlation in the Late Triassic and earliest Jurassic was to rely on the thickest and longest sedimentary sections available from both continental and marine environments that contain one or more chronostratigraphic proxies for dating and correlation. The initial framework was established from continuous scientific and geotechnical coring of ~5000 m-thick terrestrial deposits in the Newark basin under the Newark Basin Coring Project (NBCP) and Army Corps of Engineers (ACE) cores (Olsen et al., 1996a) (Fig. 1). These and related strata of the Newark Supergroup in the rift basins of Eastern North America range in age from the Carnian to the Hettangian and Sinemurian based on palynofloral zones and land vertebrate faunachrons (Cornet and Traverse, 1975; Huber et al., 1993; Olsen and Cornet, 1988; Olsen et al., 2011; Weems and Olsen, 1997). Most significantly, the lacustrine facies that can be found in all but the lower portion of the thick Newark basin sedimentary section pervasively reflects the imprint of Milankovitch climate cyclicity and this orbital signal provides a means of pacing the deposition on a much finer 10–100 kyr scale (Olsen, 1986).

The Newark and related rift basin rocks were initially regarded as recording only normal polarity (the Newark Normal Interval of Perchersky and Khramov (1973) and Graham Normal Interval of McElhinny and Burek (1971)). However, it eventually became clear that this applied only to the disproportionately studied Newark Supergroup igneous rocks whereas the much thicker sedimentary units of much longer duration beneath the lavas were soon observed to have numerous normal and reverse polarity magnetozones; the stratigraphic distribution of the magnetozones was initially assembled from basin-wide sampling of outcrops and short cores (McIntosh et al., 1985; Witte and Kent, 1989; Witte et al., 1991) and culminated with the detailed paleomagnetic and chronostratigraphic record from the NBCP cores (Kent et al., 1995; Olsen and Kent, 1996). After some refinements in the polarity sequence, the resulting astrochronostratigraphic polarity time-scale (APTS) (Kent and Olsen (1999); see Olsen et al. (2011) for

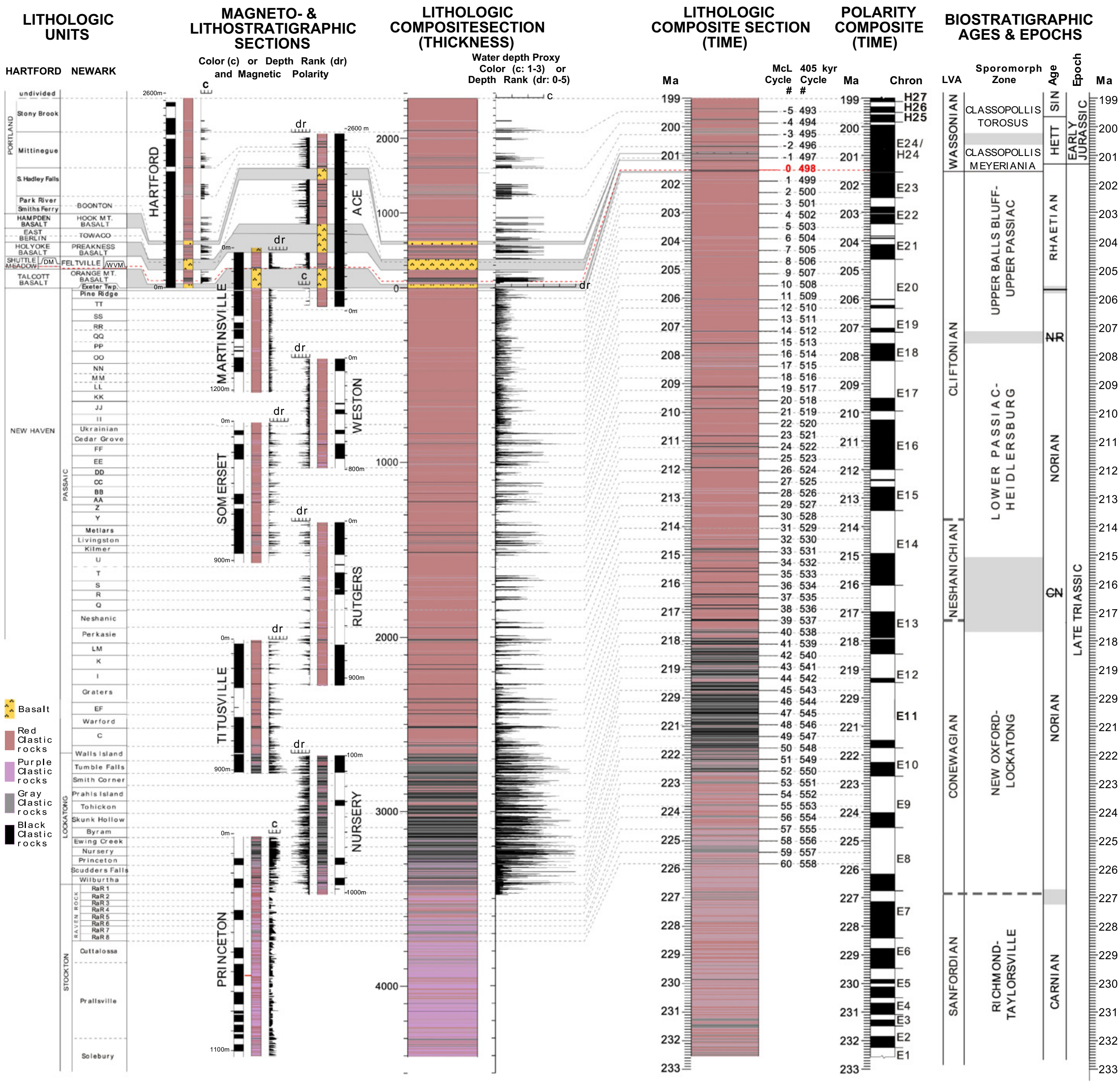


Fig. 1. Stratigraphic framework for the Newark-Hartford APTS. Lithologic Units columns show subdivisions of Hartford and Newark basin sections into formations and members (Olsen, 1980; Olsen et al., 1996a; Kent and Olsen, 2008; Whiteside et al., 2011a). Magnetostratigraphy and lithostratigraphy are based on sections from outcrop studies and several short geotechnical cores from the Hartford basin (Kent and Olsen, 2008) and in the Newark basin, outcrop studies and numerous short Army Corps of Engineers (ACE) geotechnical cores (Witte and Kent, 1990) plus seven long scientific drilling cores (Martinsville, Weston, Somerset, Rutgers, Titusville, Nursery and Princeton) from the Newark Basin Coring Project (Kent et al., 1995). Magnetic polarity is indicated by filled and open columns for normal and reverse polarity. Dominant lithologies are shown by purple hues (fluvial sandstones), mainly in the Stockton Formation but largely mudstones in other units, dark hues (gray to black lacustrine fissile mudstones and coarser units) mainly in the Lockatong Formation but interspersed with reddish hues (red lacustrine to fluvial mudstones, siltstones, and sandstones) that characterize much of the Passaic Formation and units interbedded with and overlying lavas (yellow hue) of the Central Atlantic Magmatic Province (Olsen et al., 1996a). The composite Newark-Hartford section is scaled from thickness to time by assuming the 66 McLaughlin cycles represent the 405 kyr eccentricity climate modulation anchored to 201.5 Ma at the base of the Washington Valley Member (WVM) and correlative position within the Durham Member (DM), corresponding to 201.6 Ma for the onset Chron E23r (= base of the Exeter Twp. Member). Polarity chrons with prefix E based on the Newark basin section and H for the Hartford basin section. Biostratigraphically-based ages and epochs assembled from Huber et al. (1993) for Land Vertebrate Age (LVA), Cornet and Olsen (1985) for sporomorph zones with earlier but no longer valid correlations to standard geologic age boundaries stricken through (~~NR~~ for Norian/Rhaetian, ~~CN~~ for Carnian/Norian), and Channell et al. (2003), Hüsing et al. (2014), Maron et al. (2015) and Muttoni et al. (2004a), amongst others (see Text), for the currently accepted correlations to the standard marine-based ages of the Late Triassic and Early Jurassic.

summary of various Newark APTS variants) was anchored to available U-Pb dates (rounded to 202 Ma) from the Palisade Sill tied to the oldest basalt lava sequence (Orange Mountain Basalt). This flood basalt event, now recognized as part of CAMP (Marzoli et al., 1999), was determined to have been short-lived (only about 600 kyr) based on Milankovitch cyclicity in the sedimentary strata interbedded with the lavas (Olsen et al., 1996b). The polarity and climate cycle sequences were subsequently extended upward by several million years into the Early Jurassic in the >2000 m of outcrop strata above the CAMP basalts in the Hartford basin (Kent and Olsen, 2008), which together with the data from the Newark basin, provide the working framework for the Late Triassic–Early Jurassic Newark-Hartford APTS.

With the general lack of radiometrically dated horizons or clear Milankovitch signatures in most Late Triassic marine sections, magnetostratigraphy has become a powerful tool for correlation to the Newark-Hartford APTS to develop an integrated global chronology. Workers eventually sought the thickest and longest marine sections (although not necessarily the most fossiliferous) to minimize the distorting effects of discontinuities in deposition on the magnetic polarity pattern. Some of the key reference sections with magnetobiostratigraphy for this time frame are Silicka Brezova in Slovakia (Channell et al., 2003) and Pizzo Mondello in Sicily (Muttoni et al., 2004a) for the Carnian and Norian, Brumano-Italcementi in the Southern Alps (Muttoni et al., 2010) and Pignola-Abriola in Southern Italy (Maron et al., 2015) for the Norian and Rhaetian, and St. Audrie's Bay and East Quantoxhead in Britain for the Rhaetian to Sinemurian, most of which now also has an astronomically-calibrated magnetobiostratigraphy (Hounslow et al., 2004; Hüsing et al., 2014; Ruhl et al., 2010). Available U-Pb zircon dates on ash layers in marine sections, notably from Peru (Guex et al., 2012; Schaltegger et al., 2008; Schoene et al., 2010; Wotzlaw et al., 2014) and Italy (Furin et al., 2006; Mietto et al., 2012), lack magnetostratigraphic control and depend on biostratigraphic correlation or the dates themselves for integration with the APTS. Hounslow and Muttoni (2010) provide an extensive catalog and discussion of magnetobiostratigraphic sections for the entire Triassic, with an emphasis on marine stage boundary definitions and correlations.

The plan of the paper is to 1) outline the development of the APTS framework from the polarity sequence and cycle stratigraphy of the Newark and Hartford basins, 2) anchor the APTS at the base of Chron E23r and the Exeter Township (Twp.) Member to the best-fit 405 kyr eccentricity cycle counting from the present using published high-precision U-Pb zircon dates on the earliest CAMP volcanics and available astronomical solutions, 3) review correlations of the magnetostratigraphy of key marine sections and standard marine stages to the APTS, and 4) compare and test the APTS with independent U-Pb dates from outside the Newark and Harford basins. We conclude by offering motivations and suggestions for further work and prognoses of their likely success.

2. Assembly of Newark-Hartford APTS

2.1. Newark and Hartford magnetozone sequence

For the Late Triassic, the framework for the paleomagnetic polarity sequence is constructed from data from seven partially overlapping NBCP drill cores totaling >6700 m in length with better than 99% core recovery (Table 1). The cored sections were compiled into an approximately 5000 m-thick composite section with about 25% redundancy in stratigraphic coverage (Olsen et al., 1996a). The initial sampling interval of nominally 2.75 m (one sample plug of the most favorable lithology per (9′) core box) yielded 2400 samples. Early acquired characteristic magnetizations could be isolated from hematite-bearing as well as magnetite-bearing sediment facies. Principal component analysis of progressive thermal demagnetization data allowed the delineation of 59 polarity intervals ranging from 4 m to over 300 m in stratigraphic thickness (Kent et al., 1995) (Fig. 1). Of great importance are the tests of lateral continuity and consistent relationship of the magnetozones to lake level cycles in the stratigraphically overlapping sections of the drill cores. In all six comparisons between the overlapped portions of the seven NBCP cores, some up to 42 km apart (Fig. 2), there was no violation between the lithologic marker beds (i.e., cyclostratigraphy) and polarity magnetozones, demonstrating their mutual validity as time horizons. Every overlap zone has at least one polarity transition and the polarity is consistent even when lateral facies changes include a change of color and magnetic mineralogy. This dataset was supplemented by higher density sampling (every 0.3 m or so) across 42 magnetozone boundaries representing 35 different polarity reversals in the lacustrine Lockatong and Passaic Formations (although no additional sampling was done in the fluvial Stockton Formation) (Kent and Olsen, 1999). Each pairwise correlation of the overlap zones of the core demonstrates the reproducibility of both the magnetostratigraphy and cyclostratigraphy over much of the central Newark basin.

The reverse polarity magnetozone at the base of the recovered section (Stockton Formation in the Princeton NBCP core) is designated E1r and succeeding pairs of predominantly normal and predominantly reverse polarity intervals are numbered upward as E2n, E2r; E3n, E3r; etc., with the suffix for the dominant polarity (n is normal polarity, r is reverse) of each constituent submagnetozone. To balance the stratigraphic thickness of first rank magnetozones over the entire section and to avoid having short polarity intervals at their boundaries, selected shorter polarity intervals were assigned subordinal integers (ascending upsection) that are appended after a decimal point to the higher-order magnetozone designation and also given a suffix indicating dominant polarity, for example, E22n is comprised of E22n.1n, E22n.1r, and E22n.2n. The overall nomenclature scheme is similar to that for the Late Jurassic to Cenozoic polarity sequence derived from marine magnetic anomalies (e.g., Cande and Kent, 1992; Channell et al., 1995; Harland et al., 1982) except those were numbered from youngest to oldest whereas the polarity magnetozones and corresponding chrons for the Hartford-Newark APTS are numbered from oldest to youngest in deference to standard stratigraphic practice.

In the Newark basin, the uppermost 1000 m-thick section of interbedded volcanics and continental sediments of mostly Early Jurassic age is characterized by normal polarity. This was known from early paleomagnetic studies (Opdyke, 1961) and subsequently confirmed by data from outcrop and short geotechnical cores (McIntosh et al., 1985; Witte and Kent, 1990) as well as the stratigraphically highest (Martinsville) NBCP core (Kent et al., 1995). This thick normal polarity magnetozone was elevated in rank to E24n from its previous designation E23n.2n (Kent et al., 1995) on the strength of data supporting the existence of the preceding short reverse polarity interval that became E23r (from E23n.lr) (Kent and Olsen, 1999). In the nearby Hartford basin, paleomagnetic study of samples from 80 sites and 15 geotechnical cores distributed over a 2500 m-thick section of sedimentary units that are interbedded with and overlie the same sequence of lavas as in the Newark basin extends the polarity record another 2 Myr according to cycle stratigraphic analysis (Kent and Olsen, 2008) (Fig. 3). A magnetostratigraphy was not developed in the fluviatile New Haven Arkose that underlies the lavas in the Harford basin because of a lack of marker beds to tie together the discontinuous outcrops. Nevertheless, the correlative Chron E24n in the Harford basin was designated H24n for convenience and the succeeding polarity magnetozones are labeled H24r through H27n (partim).

There are a total of 66 magnetozones in the Newark and Hartford basin sections with the lowermost (E1r) and uppermost (H27n) polarity zones are perforce unbounded stratigraphically. The thickest magnetozone at just over 1500 m is E24n/H24n, which includes all

Table 1
Corrected locations[a] for NBCP core sites.

Name	Decimal degrees		Degrees Minutes Seconds	
	Latitude North	Longitude West	Latitude North	Longitude West
Martinsville	40.611446°	74.574368°	40°36′41.21″	74°34′27.72″
Weston	40.542116°	74.562873°	40°32′31.62″	74°33′46.34″
Somerset	40.505764°	74.565386°	40°30′20.75″	74°33′55.39″
Rutgers	40.526411°	74.433083°	40°31′35.08″	74°25′59.10″
Titusville	40.318858°	74.849922°	40°19′07.89″	74°50′59.72″
Nursery	40.289598°	74.823748°	40°17′22.55″	74°49′25.49″
Princeton	40.361275°	74.613286°	40°21′40.59″	74°36′47.83″

[a] GoogleEarth Coordinates (Map Datum WGS84).

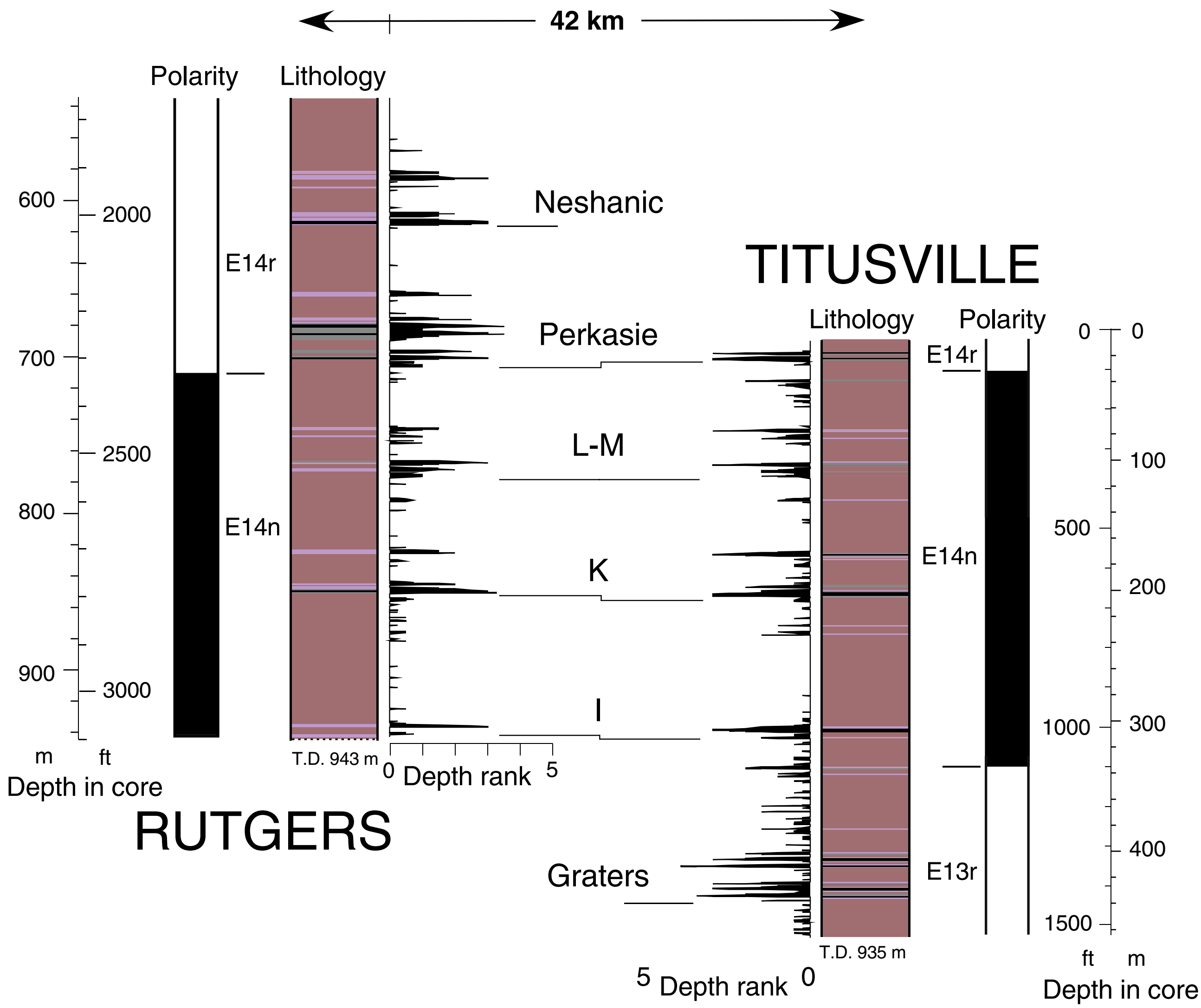


Fig. 2. An example of lateral correlation of magnetozones, lithofacies and lake depth ranks between stratigraphically overlapping NBCP drill cores, in this case from the Rutgers and Titusville sites that are separated by 42 km (Kent et al., 1995; Olsen et al., 1996a).

the CAMP lavas (Prevot and McWilliams, 1989); the next thickest ones are H25n (380 m) in the Portland Formation and E11r (304 m) m in the Lockatong Formation. At the opposite end of the spectrum, there are a number of magnetozones only a few meters thick: E22n.1r (1.6 m), E21r.2n (2.8 m), and notably E23r (3.1 m). The overall mean magnetozone thickness is 104 m (in Rutgers-normalized depth units for the Newark basin).

2.2. Newark and Hartford orbital climate cycles

The Newark, Hartford and other Eastern North American rift basins were occupied by very large playa to perennial lakes during their long histories (Olsen, 1990). Lacustrine deposits characterize virtually all but the basal (fluvial) Stockton Formation of the Newark basin including the sedimentary units interbedded with and overlying the CAMP volcanics in both the Newark and Hartford basins, with the exception of the uppermost strata of the latter. These lacustrine sediments display a pronounced cyclic variation in lithofacies (Olsen, 1986; Van Houten, 1964). The cyclicity can be characterized by facies or depth ranks, a measure of lake depth based on a classification of water-depth-related sedimentary facies suitable for numerical analysis (Olsen, 1986; Olsen and Kent, 1996). To a less subtle extent the cyclicity is also evident in the color of the strata which reflects redox conditions of the sediments and other physical lithological characteristic as seen in instrumental natural gamma ray and sonic geophysical logs (Goldberg et al., 1994; Olsen and Kent, 1999; Reynolds, 1993).

The fundamental lithofacies variation is the Van Houten cycle, which is recognized on a stratigraphic scale of 3 to 6 m in the NBCP cores (Fig. 4). This cycle was hypothesized (Van Houten, 1964) and confirmed (Olsen, 1986; Olsen and Kent, 1996) to correspond to lake level (i.e., climate) change at a precessional periodicity (nominal 20 kyr, modeled as roughly 5% shorter than today's 21 kyr average period because of recession of the Moon; Berger et al., 1992). The expression of Van Houten cyclicity is modulated by several orders of orbital eccentricity variations, most prominently by the 405 kyr long eccentricity orbital variation expressed as the McLaughlin cycle (Olsen, 1986; Olsen and Kent, 1996) as well as the short and very long eccentricity variations of ~100 kyr and ~1.8 Myr.

The cyclicity of varying amplitudes and periods permeates the entire lacustrine sedimentary section as evident in moving window spectral analysis (Olsen and Kent, 1999) and wavelet analysis (Fig. 5) of depth ranks even with no tuning. The multiple periodicities present change in concert with the specific thickness of the McLaughlin cycle, which is prima facie evidence of pacing by an external clock and otherwise inexplicable. Furthermore, periodic fluctuations in amplitude in higher frequencies are evident in the band of power at the appropriate lower

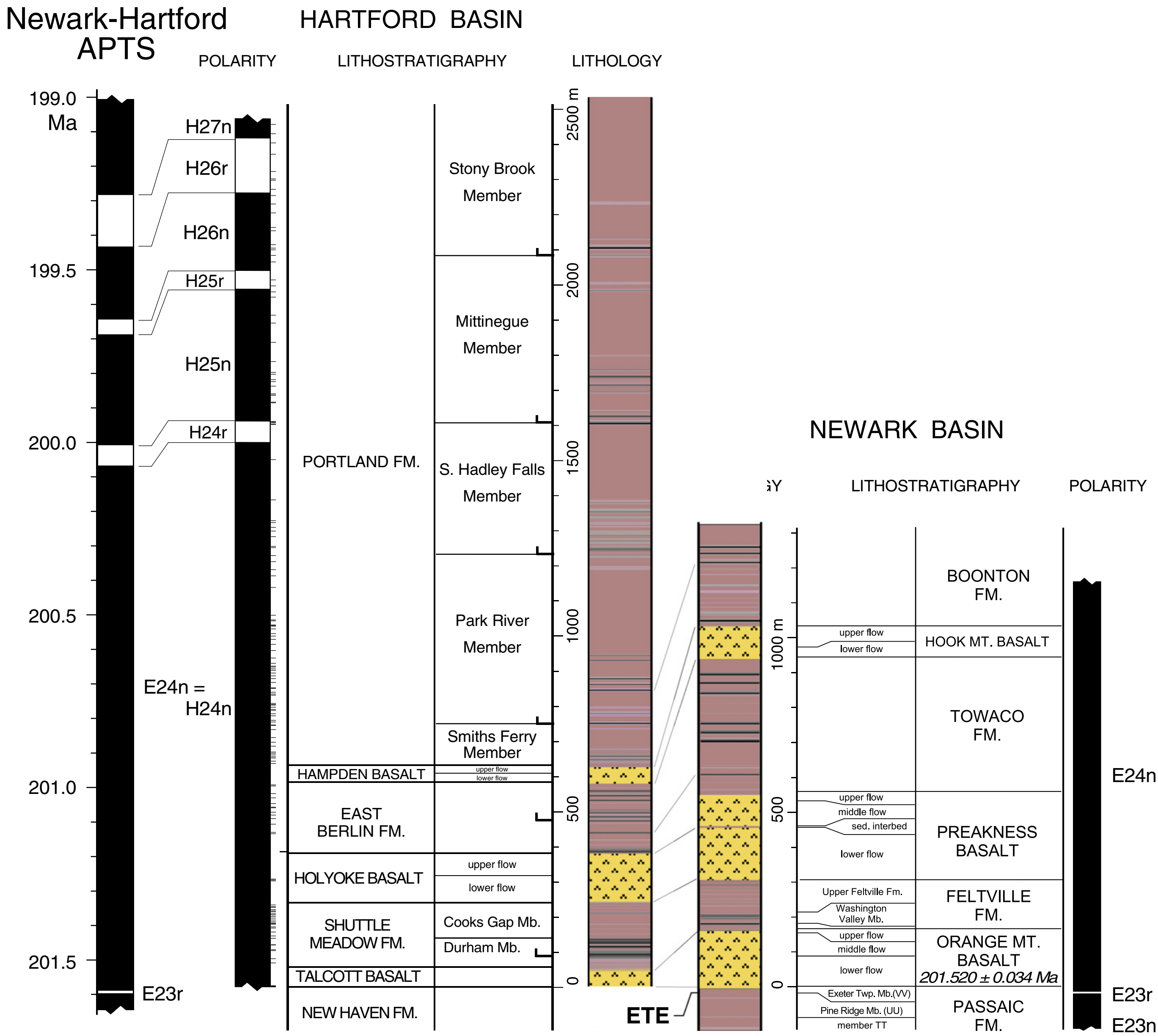


Fig. 3. Magnetostratigraphy and lithostratigraphy of latest Triassic and Early Jurassic strata including lava units of the Central Atlantic Magmatic Province in the Hartford basin compared to the correlative section from the Newark basin. Filled and open bars are for normal and reverse polarity in the polarity columns, which for the Hartford basin also shows the stratigraphic sample levels by ticks on the right edge. The 405 kyr McLaughlin cycles in the Hartford basin, as indicated by heavy lines with upturned line segments in the lithostratigraphy member column, were used to estimate ages anchored to 201.5 Ma for base of black shales in the Durham Member for the Early Jurassic portion of the composite Newark-Hartford APTS. Figure is updated from Kent and Olsen (2008).

frequencies. There is strong consistent power with a period of around 60 m that is present throughout, even if it is faint in the upper 800 m of the section where the depth ranks are muted. This is the McLaughlin cycle, which is a robust lithologic variation that effectively corresponds to mappable lithostratigraphic members in the Lockatong Formation and most of the Passaic Formation in the Newark basin as well as most of the Portland Formation of the Hartford Basin. This practice follows the basic pattern laid down by D. B. McLaughlin (McLaughlin, 1943, 1945, 1959) for which the cycle is named. Simply setting the duration of the Upper Stockton, Lockatong and Passaic Formations as most probably longer than 10 Myr and <30 Myr makes it possible to associate the spectral peak at around 60 m wavelength with the long (405 kyr) orbital eccentricity cycle.

There is notable power at other periodicities as well (e.g., at around 20 m) such that the overall spectrum of frequency variations makes a very compelling case for Milankovitch climate forcing. For example, a comparison of the average power spectrum of depth ranks for the Lockatong Formation to the spectrum of expected climate forcing in the tropics from modern orbital changes (Laskar et al., 2011) shows a remarkable parallel in the family of eccentricity peaks (twin peaks at around 100 kyr, and a single peak at around 400 kyr) (Fig. 6). This pattern supports the interpretation of associating the McLaughtlin cycle with the 405 kyr eccentricity variation, the largest and most stable term in the approximation of eccentricity of Earth's orbital variations on geologic time-scales (Berger et al., 1992; Laskar, 1990; Laskar et al., 2004, 2011). The virtual absence of any power that would correspond to an obliquity cycle is noteworthy and greatly simplifies the interpretation of the pattern of climate forcing in these sediments.

We use the 405 kyr McLaughlin cycle to generate a relative chronology for the Late Triassic and earliest Jurassic polarity sequence from the

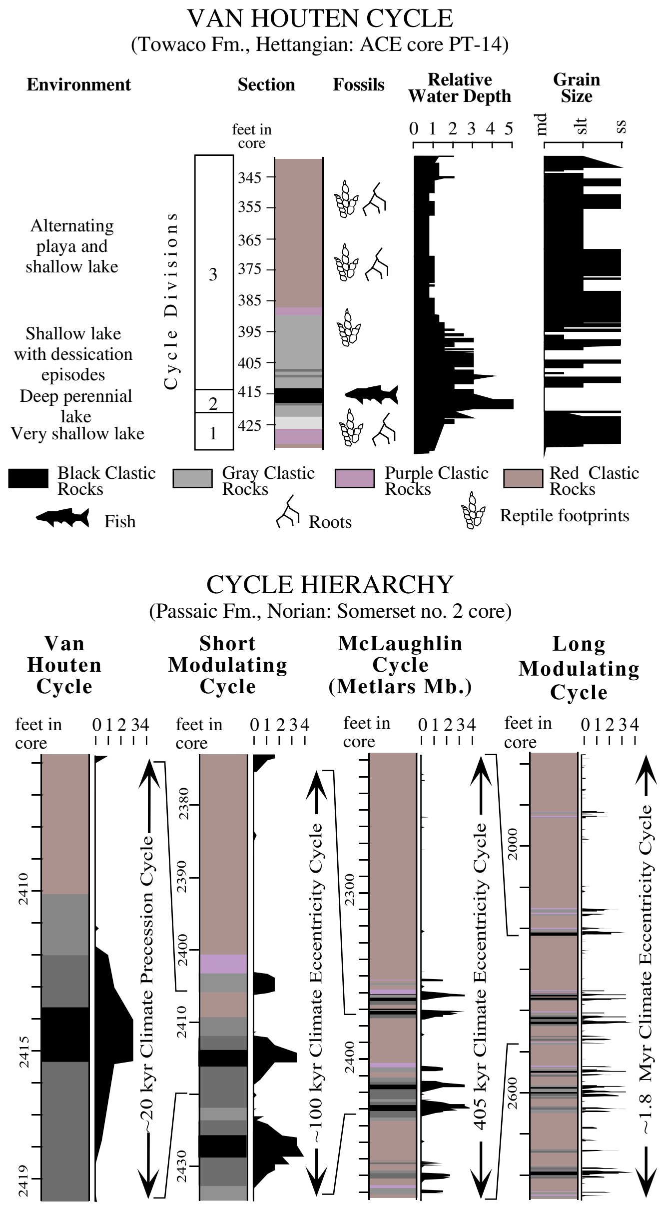


Fig. 4. Van Houten and longer modulating lacustrine cycle types of the Newark basin. A. The fundamental Van Houten cycle (example based on section in ACE core PT-14 in New Jersey). B. Van Houten cycle and the nested short, McLaughlin, and long modulating cycles of the lacustrine rocks of the Newark basin illustrating the method of member delineation using as an example the Metlars Member of the Passaic Formation as seen in the NBCP Somerset no. 2 core. Figure adapted from Olsen et al. (1996a, 1996b).

Newark and Hartford basins. In the uppermost Stockton, Lockatong and Passaic formations of the Newark basin, there are 60 McLaughlin cycles averaging 62 m in thickness (Kent and Olsen, 1999; Olsen and Kent, 1996). Coinciding with and following the emplacement of the CAMP volcanics, McLaughlin cycles become much thicker, on the order of 400 m. This thickening is seen in the Feltville and Towaco Formations of the Newark basin and the Shuttle Meadow and East Berlin Formations of the Hartford Basin interbedded with the volcanics, as well as in the Boonton and Portland formations overlying the extrusive intervals in the Newark and Hartford basins, respectively (Kent and Olsen, 2008).

McLaughlin cycles are lithological units with boundaries defined by specific beds. The convention employed in designating them is to place the base of the McLaughlin cycle at the first well-developed and mappable darker-colored bed (generally gray or black) producing an asymmetrical-looking cycle (Olsen et al., 1996a) (Fig. 4VH). As a consequence, the base of each McLaughlin cycle does not conform exactly to the peak of a 405 kyr cycle, when the amplitude of lake depth changes from precessional variations would be the greatest, but rather tends to be offset below it by two or more precessional cycles. However, the effect is not cumulative and there is essentially no net effect over multiple cycles in the chronology because the offsets tend to be systematic.

2.3. Astrochronological age model anchored by U-Pb dates

As a framework for a global geologic and polarity time-scale, the astrochronology of the Newark-Hartford composite sequence needs to be anchored to a datum of known age in millions of years before present. The best-dated levels that can be directly tied to this magnetic reversal and orbital sequence are high-precision U-Pb zircon dates from CAMP rocks in the Newark basin (Blackburn et al., 2013). Of particular utility is the date of 201.520 ± 0.034 Ma for the Palisade Sill, which is linked to the Orange Mountain Basalt, the oldest series of CAMP lavas in the Newark basin. A thin reverse polarity magnetozone (E23n.1r) was found about 20 m below the Orange Mountain Basalt in the initial sampling of the Martinsville NBCP core (Kent et al., 1995) and soon afterwards confirmed >100 km away in the fossiliferous Grist Mill section of the Jacksonwald Syncline (Fig. 27 in Olsen et al. (1996a)), where it had been narrowly missed between outcrop sites TJJ and TJR of Witte et al. (1991) but was subsequently identified at Grist Mills and the adjacent Exeter section about 1 km to the east (Olsen et al., 2002b). Elevated in rank and renamed E23r, the reverse polarity magnetozone was more precisely delineated in the Martinsville NBCP core with higher density sampling (Kent and Olsen, 1999).

To register these U-Pb dated basalt flows and associated magnetic polarity zones into the 405 kyr McLaughlin-cycle-based astrochronology, it is necessary to use a finer-scale approach to the astrochronology of the lacustrine strata interbedded with and surrounding the basalt units because the counterparts to McLaughlin cycles are themselves segmented by the flows. It has been clear since the 1980s (e.g., Olsen, 1980, 1986) that the overall hierarchy of cycles seen in the Passaic Formation continues into the overlying strata and is in fact amplified to a remarkable extent.

At the largest scale, Van Houten cycles with the best-developed microlaminated deeper-water units (with abundant articulated fossil fish) occur in three intervals in the sedimentary strata bounding and interbedded with the CAMP lava flows. The lowest interval is in the Washington Valley Member of the Feltville Formation. The next higher interval is in the middle Towaco Formation, and the uppermost interval with well-developed Van Houten cycles is in the Boonton Formation as seen in the ACE cores. These deeper-water units have been interpreted as occurring at times of maximum precessional variability and hence at the peaks in eccentricity of the ~100 kyr cycle coinciding with the peak of the 405 kyr cyclicity (Olsen and Kent, 1996). Other clusters of laminated dark mudstones in Van Houten cycles occur on both sides of each of these peaks and demarcate lesser peaks in the ~100 kyr cycles. Intervals of very poorly developed Van Houten cycles nearly lacking gray or black beds occur in the lower, but not lowest Boonton Formation, the upper, but not uppermost Feltville Formation and just below the uppermost few tens of meters of the Passaic Formation. These intervals correspond to troughs in precessional variability and hence troughs in the 405 kyr eccentricity cycle.

The lowest McLaughlin cycle associated with the CAMP is thus split by the Orange Mountain Basalt into a lower part with the underlying Exeter Twp. Member (=Member VV) and an upper part with the overlying Washington Valley Member, making these two members conceptually different from the lithostratigraphic members of the rest of the Passaic Formation. Likewise, the next McLaughlin cycle has the Preakness Basalt nearly at where its base would be otherwise drawn, and the Hook Mountain Basalt occurs nearly exactly at the top of the same McLaughlin cycle. This approach has been used consistently for

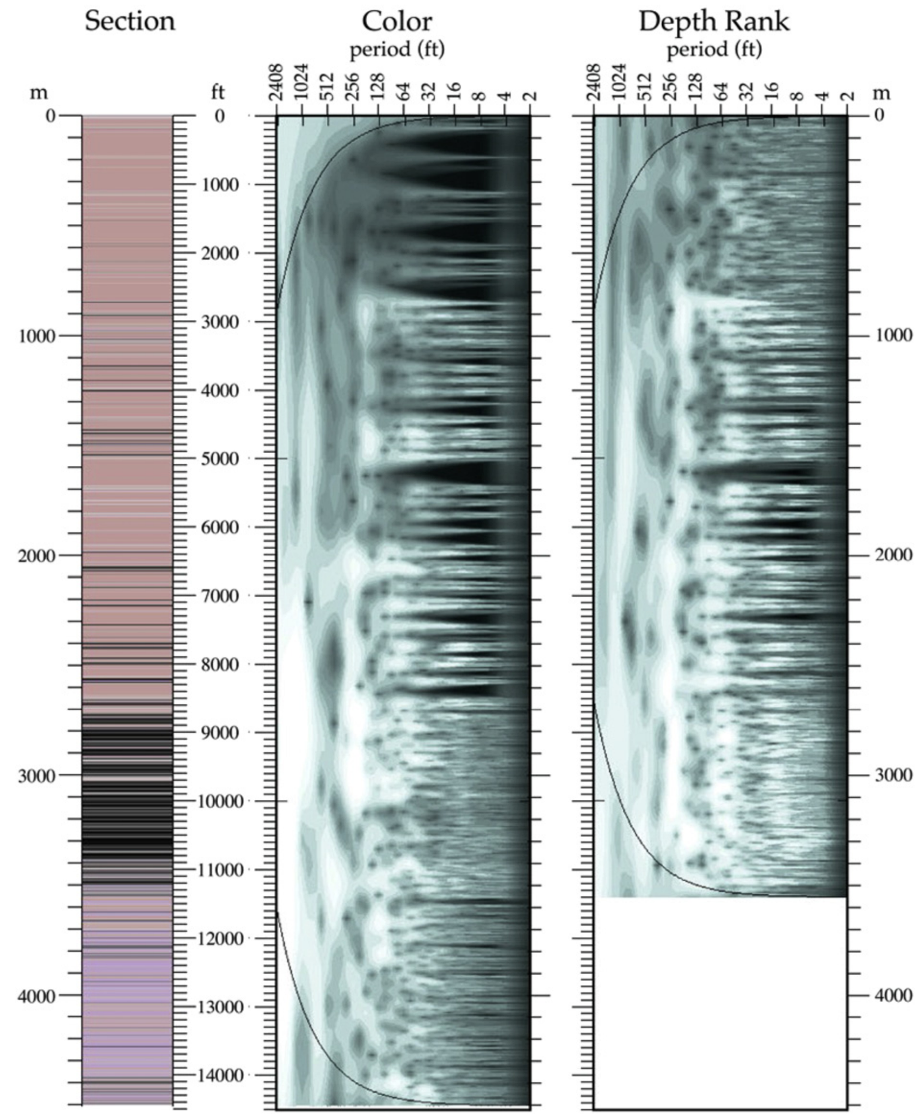


Fig. 5. Continuous wavelet transform proxies of relative lake-depth (color and depth ranks) for composite section of NBCP cores in untuned (i.e., common depth) scale. White is high and black is low power. Note consistent structure of spectrum from base of the lacustrine sequence at about 3900 m (base Raven Rock Member of Stockton Fm.); major (vertical) bands of power are at very approximately 8, 64, 256, and 1024 ft., that correspond to the precession (~20 kyr), short eccentricity (~100 kyr), long-eccentricity (405 kyr) and the g3-g4 eccentricity (1.8 Myr) bands, respectively. Lateral shifts in these bands occur in unison at ~3200 m and ~1900 m, most simply interpreted as accumulation rate changes. The constancy of the ratios between thickness periods (vertical bands of high power), independent of average accumulation rate is inexplicable without extrinsic (i.e., Milankovitch) climate control at multiple simultaneous periods. Analyses were conducted using Matlab wavelet script provided by C. Torrence and G. Compo (http://paos.colorado.edu/research/wavelets/).

the last 20 years for the CAMP-related deposits and to estimate the duration of the CAMP events (Blackburn et al., 2013; Kent and Olsen, 2008; Olsen et al., 1996b, 2003; Whiteside et al., 2007).

All of the variation amongst the Van Houten cycles seen in the strata around and in the CAMP lava flow interval fits relatively easily within two successive 405 kyr cycles, providing the basis for a 20 kyr-level astrochronology (Olsen et al., 1996b; Whiteside et al., 2007) that was tested and well corroborated by high-precision zircon U-Pb zircon dates by Blackburn et al. (2013). Given this astrochronology, with the peaks of the 405 kyr cycles pegged at the Washington Valley Member and the middle Towaco Formation, the rest of the Newark basin astrochronology can be calibrated and the magnetostratigraphy directly linked with it, as follows, to derive APTS.

At the Exeter outcrop section in the Jacksonwald Syncline, which contains rich litho-, chemo-, and biostratigraphic signals of environmental changes (Olsen et al., 2002b, 2011; Whiteside et al., 2010), the base of E23r is found to be closely aligned with the base of the Exeter Twp. Member (Fig. 7). Chron E23r encompasses about ½ (~11 kyr) of a Van Houten precession cycle between the base of the Exeter Twp. Member and a gray mudstone containing evidence from sporomorphs for the ETE, which at Jacksonwald is estimated to occur 14 kyr before the Orange Mountain Basalt or 20 kyr after the beginning of Chron E23r. The onset of Chron E23r is thus estimated from astrochronology to be 34 kyr before the Orange Mountain Basalt. Although conveniently situated as a mappable unit with respect to Chron E23r, the ETE, and the Orange Mountain Basalt, the Exeter Twp. Member, as described above, is actually not the next McLaughlin cycle after the Park Ridge Member (= member UU); that role is assigned to the Washington Valley Member whose base is equivalent to about one short eccentricity cycle above the base of the Exeter Twp. Member. In terms of calibration (with the ages to the nearest 0.10 Myr), the base of the Washington Valley Member can thus be set to 201.50 Ma, making the base of the Exeter Twp. Member and coincident Chron E23r at 201.60 Ma.

If the 405 kyr orbital cycle is indeed stable over geological timescales and the U-Pb geochronology is accurate within analytical uncertainties, the base of the Washington Valley Member, a gray and black thin-bedded mudstone corresponding to relatively high lake level and picked to coincide close to a local long eccentricity (Ecc405) maximum,

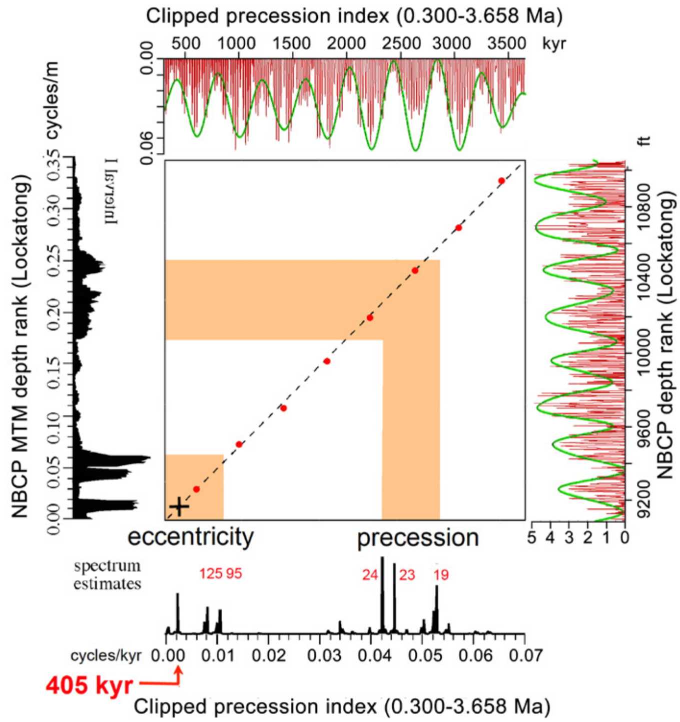


Fig. 6. Comparison of power spectra from clipped precession index for past 3.5 Myr (x-axis) with that of a comparable duration of depth ranks for the Lockatong Formation of NBCP section on thickness scale (y-axis) showing rationale for assigning spectral peaks in Newark record to those from orbital climate forcing. The cross marks the 405 kyr eccentricity peak used for construction of a time-scale.

should be in phase with an eccentricity maximum of an Ecc405 cycle projected back by the nearest integer number from the most recent calculated eccentricity maximum. An eccentricity maximum ($k = 1$) last occurred at 0.216 Ma (Laskar et al., 2004) (Fig. 8). The base of the Washington Valley Member should therefore have occurred in the maximum eccentricity phase of cycle Ecc405:k, where

$$(k-1) = (201.5-0.216)/0.405; k = 497.998 \quad (1)$$

This estimate of k is remarkably close, indeed virtually identical to an integer value expected for being in the same phase of long eccentricity counting down from the present. Departures from an integer value of k for the Washington Valley Member could occur for any number of reasons, from uncertainties of the U-Pb dating (±0.034 Myr at 95% confidence), uncertainties in estimates of sedimentation and lava extrusion rates used to extrapolate the U-Pb date to the base of the overlying Washington Valley Member, or the possibility that the laminated units used to delineate the base of the Washington Valley Member occurred one or more precession cycles from the peak eccentricity. There may even be a very small change in the Ecc405 period. However, the close correspondence in phase of the U-Pb-derived date and that calculated for the Ecc405 cycle at the Washington Valley Member suggests that

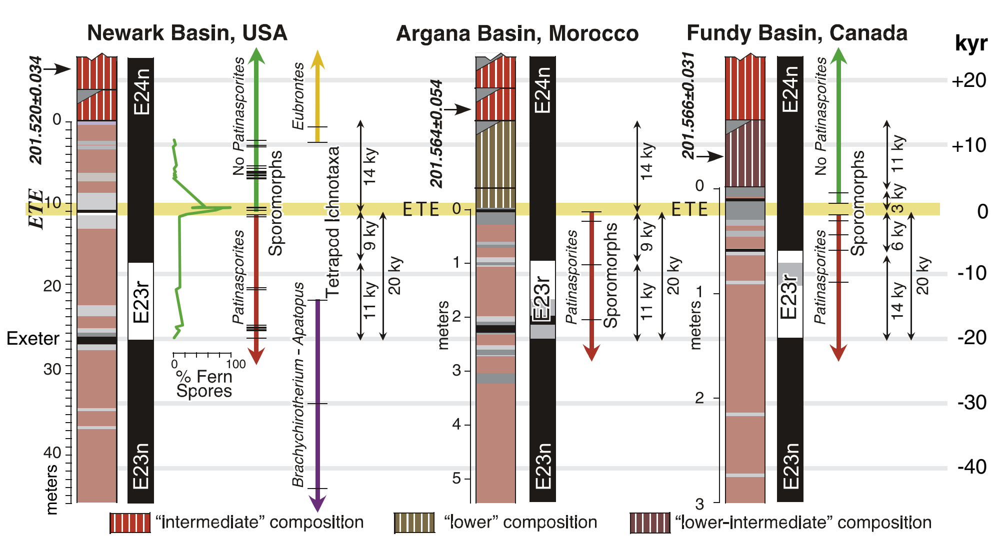


Fig. 7. Geochronologic, magnetostratigraphic, palynologic and tetrapod ichnofauna data sets for the stratigraphic interval from E23r to the lowermost CAMP basalts in the Newark, Argana, and Fundy basins (from Blackburn et al., 2013). See also exchanges on false calls of Chron E23r in lava sections in High Atlas of Morocco lava (Font et al., 2011; Knight et al., 2004; Marzoli et al., 2004, 2008; Whiteside et al., 2007, 2008).

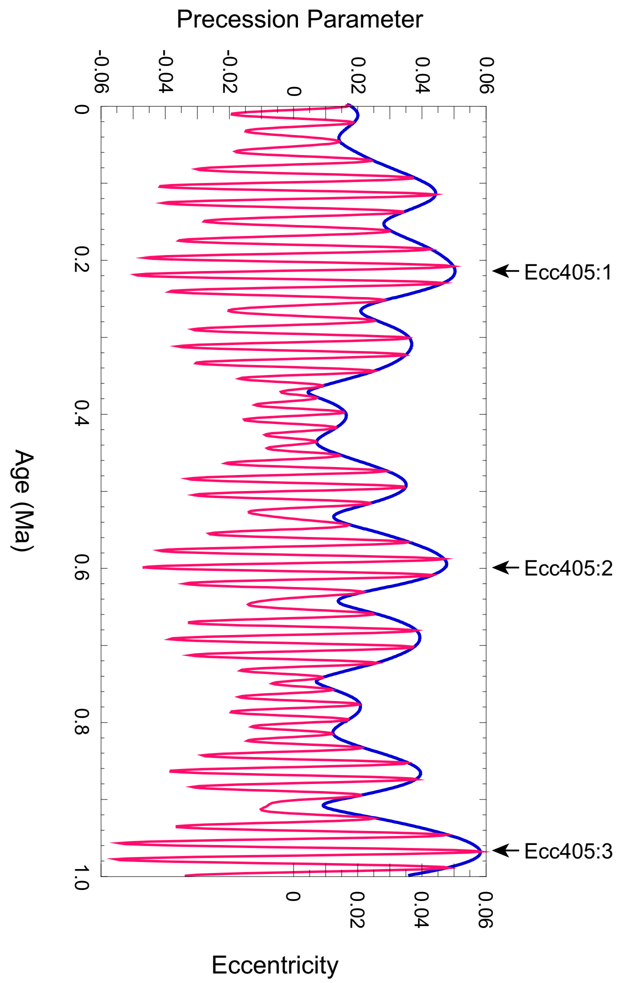


Fig. 8. Eccentricity and precession parameter for past 1000 kyr (Paillard et al., 1996) showing successively older 405 kyr eccentricity maxima (Ecc405:1, Ecc405:2, Ecc405:3, etc.). The nomenclature (Ecc405:k) is extended back into the Triassic assuming the period stays constant at 405 kyr.

the total uncertainty counting back from the present should amount to no more than plus or minus only one full 405 kyr, within the uncertainty of <0.2% cited for the Mesozoic Era (Laskar et al., 2004).

Table 2
Lithologic members as McLaughlin 405 kyr eccentricity cycles in the Newark-Hartford succession for the Late Triassic–Early Jurassic.

Member	Level (m)	McL	Ecc405 (k)	Age (Ma)
Stony Brook	2081.8	−5	493	199.48
Mittinegue	1602.9	−4	494	199.88
S. Hadley Falls	1228.5	−3	495	200.29
Park River	747.2	−2	496	200.69
Upper East Berlin + Smiths Ferry	504.8	−1	497	201.10
Shuttle Meadow + Lower East Berlin/ Washington Valley	81.0	0	498	**201.50**
Orange Mountain Basalt	*0.0*	*0.16*	*498.16*	*201.57*
ETE	*−5.2*	*0.19*	*498.20*	*201.58*
Exeter Twp. (VV)	*−12.3*	*0.25*	*498.25*	*201.60*
Pine Ridge (UU)	−59.4	1	499	201.91
TT	−124.7	2	500	202.31
SS	−186.5	3	501	202.72
RR	−241.3	4	502	203.12
QQ	−307.0	5	503	203.53
PP	−360.1	6	504	203.93
OO	−429.8	7	505	204.34
NN	−489.1	8	506	204.74
MM	−543.8	9	507	205.15
LL	−600.7	10	508	205.55
KK	−651.0	11	509	205.96
Flemington (JJ)	−728.3	12	510	206.36
II	−785.7	13	511	206.77
Ukrainian (HH)	−840.6	14	512	207.17
Cedar Grove (GG)	−898.9	15	513	207.58
FF	−969.6	16	514	207.98
EE	−1034.2	17	515	208.39
DD	−1074.9	18	516	208.79
CC	−1140.6	19	517	209.20
BB	−1199.3	20	518	209.60
AA	−1244.6	21	519	210.01
Z	−1285.7	22	520	210.41
Y	−1362.5	23	521	210.82
Metlars (Z)	−1419.8	24	522	211.22
Livingston (Y)	−1478.3	25	523	211.63
Kilmer (X)	−1518.9	26	524	212.03
U	−1598.6	27	525	212.44
T	−1677.8	28	526	212.84
S	−1730.8	29	527	213.25
R	−1785.7	30	528	213.65
Q	−1847.1	31	529	214.06
Neshanic (P)	−1947.9	32	530	214.46
Perkasie	−2033.4	33	531	214.87
L-M	−2105.4	34	532	215.27
K	−2185.2	35	533	215.68
I	−2270.2	36	534	216.08
Graters	−2375.1	37	535	216.49
E-F	−2443.5	38	536	216.89
Warford Brook	−2522.7	39	537	217.30
C	−2623.2	40	538	217.70
Walls Island	−2708.2	41	539	218.11
Tumble Falls	−2781.1	42	540	218.51
Smiths Corner	−2860.4	43	541	218.92
Prahls Island	−2938.9	44	542	219.32
Tohicken	−3007.4	45	543	219.73
Skunk Hollow	−3091.6	46	544	220.13
Byram	−3147.5	47	545	220.54
Ewing Creek	−3203.4	48	546	220.94
Nursery	−3251.0	49	547	221.35
Princeton	−3304.6	50	548	221.75
Scudders Falls	−3368.3	51	549	222.16
Wilburtha	−3415.9	52	550	222.56
RaR-1	−3463.2	53	551	222.97
RaR-2	−3506.9	54	552	223.37
RaR-3	−3542.6	55	553	223.78
RaR-4	−3583.1	56	554	224.18
RaR-5	−3624.8	57	555	224.59
RaR-6	−3654.5	58	556	224.99
RaR-7	−3698.1	59	557	225.40
RaR-8	−3739.1	60	558	225.80
Stockton lowest	−4400.9			232.57

Notes to Table 2

Stratigraphic levels (arbitrarily made negative for Newark basin where scaled to the Rutgers NBCP core and positive in Hartford basin) measured with respect to the base of the Orange Mountain Basalt (and equivalent Talcott Basalt) of the bases of the lithologic members and formations and McLaughlin (McL) cycles. Members that represent McLaughlin cycles are numbered from 0 for the Shuttle Meadow + East Berlin/ Washington Valley Member downward to +60 for the lowest Raven Rock member (RaR-8) of the Stockton Formation in the Newark basin and upward to -5 for the Stony Brook Member of the Portland Formation in the Hartford basin. McLaughlin cycles are assumed to be manifestations of the long (405 kyr) orbital eccentricity cycle; the defined base of each member as McLaughlin cycle is assumed to be in more or less consistent phase with a peak of the 405 kyr orbital eccentricity cycle (Ecc405), which are counted back (k) from the most recent peak (k = 1) at 0.216 Ma (Fig. 8). Age (Ma) = 0.216 + (k − 1) ∗ 0.405; age of lowest recovered Stockton Formation in Princeton NBCP core is estimated by extrapolation of sediment accumulation rate for RaR-1 to RaR-8. Values for stratigraphic levels in the Hartford basin have been modified from Kent and Olsen (2008) by a section added to the East Berlin Formation from cores BD 227A and 255 from Steinen et al. (2015), which added a net 34.07 m to the basal East Berlin and overlying section (see Fig. 3) but with no change to the astrochronological ages of member boundaries. Age in bold (201.5 Ma) is calibration age for APTS assigned to base of Washington Valley Member (Ecc405:498.0) from high-precision U-Pb zircon date of immediately underlying Orange Mountain Basalt, whose age along with those for ETE and Exeter Twp. Member are shown in italics because these ages are interpolated within a McLaughlin cycle (Ecc405:498).

We elect to maintain a period of 405 kyr for Ecc405 and assign an age of 201.5 Ma for the base of the Washington Valley Member as Ecc405:498 and an age of 201.6 Ma for the base of Chron E23r. We believe that our astrochronological age estimate for the base of the Washington Valley Member and the inferred age of Chron E23r do not violate any age constraints within their probable estimates of uncertainty determined by Blackburn et al. (2013). The previously identified McLaughlin cycles in the tabulation of Kent and Olsen (1999) are renumbered such that McL #60 is RaR-8 in the Raven Rock, which becomes Ecc405:558, McL #59 is RaR-7 in the Raven Rock, which becomes Ecc405:557, and so forth up to McL #0, the Washington Valley Member, which becomes Ecc405:498 from the Newark basin. The renumbering continues in the Early Jurassic of the Hartford basin (Kent and Olsen, 2008) from the Upper Berlin Fm + Smiths Ferry Member (Ecc405:497), Park River Member (Ecc405:496), and up to the Stony Brook Member (Ecc405:493).

The Ecc405:k designations are converted to time in millions of years ago according to:

$$\text{Age (Ma)} = (k-1) * 0.405 + 0.216 \tag{2}$$

These ages are listed in Table 2.

A plot of cumulative member thickness versus age shows a generally smooth progression of slopes (i.e., sediment accumulation rates) through the Triassic part of the Newark basin section (Fig. 9). In the upper part of the composite section, however, there is an abrupt increase in sediment accumulation rate by nearly an order of magnitude at around the emplacement of CAMP volcanics. Accumulation rates are from 100 to 150 m/Myr in the Late Triassic in the Newark basin and increase to around 1000 m/Myr in the Early Jurassic of the Hartford basin and even higher in the thicker, stratigraphically overlapping portion of the Newark basin. Assuming that accumulation rates were linear within each 405 kyr McLaughlin cycle, the ages of magnetozone boundaries from E8r to H27n (Kent and Olsen, 1999) (Kent and Olsen, 2008) were interpolated accordingly. In the absence of discernible cyclicity

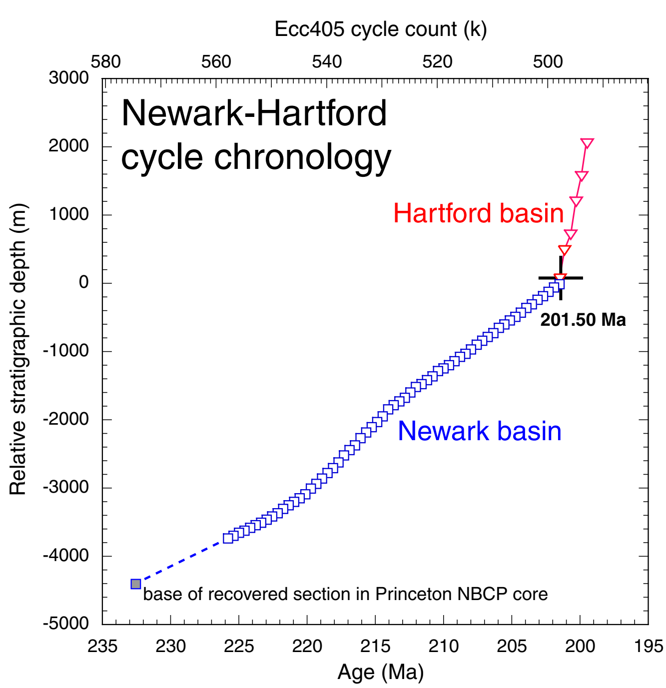


Fig. 9. Depth-age model for the Newark basin and the extrusive zone of the Hartford basin based on using 405 kyr for lithologic members recognized as McLaughlin cycles and anchored to an age of 201.50 Ma for the base of the Washington Valley Member of the Feltville Formation, just above the first CAMP lava unit (Orange Mountain Basalt) in the Newark basin.

in the more fluvial facies of the lower part of the Stockton Formation, the average sediment accumulation rate (97.7 m/Myr) for the 8 cycles in the Rock Raven unit in the upper part of the Stockton Formation was used to construct an age model for the subjacent 700 m of Newark basin section and to extrapolate ages of magnetozones E1r to E8r.

The complete APTS extends from 232.57 Ma (oldest extrapolated portion of E1r at the base of the recovered section in the Princeton NBCP core) to 199.15 Ma (youngest extrapolated portion of H27n in the Hartford basin), encompassing 33.42 Myr of record (Fig. 1). The 66 polarity interval lengths are exponentially distributed, ranging from 0.023 Myr (Chron E23r) to 1.6 Myr (Chron E24n) with an average duration of 0.51 Myr. A listing for the Newark-Hartford APTS is given in Table 3.

2.4. Comparison of Newark-Hartford APTS with previous time-scales

Olsen et al. (2011) summarized the evolution of the Newark APTS from 1995 to 2010. Beside the addition of the Jurassic chrons (H24 to H27) from the Hartford basin (as described above), principal changes from the initial Newark APTS were the addition of McLaughlin cycles RaR-1 to 8 in the upper Stockton Formation and updated numerical values of the age calibration for CAMP rocks and how it was used to anchor the relative sequence. A previous tabulation of Newark magnetochrons (Kent and Olsen, 1999) listed 202.048 Ma for the base of Chron E23r, based on available dates for the Palisade Sill that were used to project a rounded age estimate of 202 Ma for the Triassic-Jurassic boundary (at the time thought to be the same as the end-Triassic extinction, ETE) just below the Orange Mountain Basalt (as described above). The same datum was used to incorporate the Hartford magnetochrons (Kent and Olsen, 2008). Since then, U-Pb zircon dating of the North Mountain Basalt in the Fundy basin (Schoene et al., 2010) pointed to an age significantly younger than 202 Ma for the onset of CAMP volcanism. This geochronological work culminated with the comprehensive U-Pb zircon dating of CAMP rocks that confirms an age closer to 201.60 Ma for the onset of CAMP volcanism (Blackburn et al., 2013). Here we used 201.50 Ma for the base of the Washington Valley Member (Ecc405:498) with an age of 201.60 Ma (compared to 202.048 Ma) for the onset of Chron E23r (and the base of the Exeter Twp. Member) to anchor the APTS. The updated Newark-Hartford APTS is thus systematically shifted younger by about 0.5 Myr compared to some earlier versions of the time-scale (e.g., Kent and Olsen, 1999). Moreover, biostratigraphic levels like the TJB and the ETE are now free to be dated by the U-Pb calibrated astrochronology (e.g., Blackburn et al., 2013) rather than setting the timing. For example, the Geological Time Scale 2012 (Gradstein et al., 2012) and the more recent (v. 2016/12) IUGS/ISC chronostratigraphic chart (www.stratigraphy.org) set the Triassic/Jurassic boundary at 201.3 ± 0.2 Ma.

The revised Newark-Hartford APTS also eliminates an assumed 1.5 Myr gap in the record of Chron E7r that was inserted into some depictions of the time-scale (e.g., see Olsen et al., 2011). The alleged gap was based on an admittedly weak lithological correlation of cores within the Taylorsville basin and correlation of the Taylorsville composite section with the Newark basin (LeTourneau, 2003) and thought to reflect a hinge margin unconformity between tectonostratigraphic sequences II and III as seen in the Argana Basin of Morocco (Olsen, 1997). However, more recent work in the Dan River basin (Olsen et al., 2015) shows a good cross correlation with the Newark section with no gap in E7r suggesting that the Newark and Dan River basins records preserve the correlative conformity of the sequence boundary, should it exist, and that the composite correlation in the Taylorsville basin is in error. The duration of Chron E7r can be assessed independently of the Newark using the cyclicity of the Cumnock facies in the Dan River basin (Olsen et al., 2015) and the result is <400 kyr in all three cores that were studied. This is consistent with the extrapolation in the Princeton NBCP core from the Raven Rock Members in the Stockton Formation to Chron E7r. The original interpretation of the polarity sequence in the Dan River basin (Kent and Olsen, 1997) holds up completely and any change in the accumulation rate in the Stockton Formation must be below Chron E7r. Additional paleomagnetic data with cycle stratigraphic control would be useful to confirm and refine the polarity pattern for the basal fluvial part of the Stockton Formation in the Newark section, which is presently known only from the lower part of the Princeton NBCP core.

Table 3
Newark-Hartford APTS constrained by U-Pb zircon ages for CAMP.

Chron	Level (m)	McL	Ecc405 (k)	Age (Ma)
H27n	2421.0	−5.81	492.19	199.15
H26r	2266.6	−5.44	492.56	199.30
H26n	2042.7	−4.92	493.08	199.51
H25r	1991.6	−4.81	493.19	199.55
H25n	1613.0	−4.02	493.98	199.87
H24r	1554.1	−3.87	494.13	199.93
OMB	*0.0*	*0.16*	*498.16*	*201.57*
H24n = E24n	−10.7	0.22	498.22	201.59
E23r	−13.8	0.25	498.25	201.60
E23n	−152.4	2.45	500.45	202.49
E22r	−197.7	3.20	501.20	202.80
E22n.2n	−230.9	3.78	501.79	203.03
E22n.1r	−232.6	3.81	501.81	203.04
E22n.1n	−288.4	4.72	502.72	203.41
E21r.3r	−333.3	5.49	503.49	203.73
E21r.2n	−336.1	5.55	503.55	203.75
E21r.2r	−353.6	5.88	503.88	203.88
E21r.1n	−359.9	6.00	504.00	203.93
E21r.1r	−392.4	6.46	504.46	204.12
E21n	−476.6	7.78	505.77	204.65
E20r.2r	−665.2	11.18	509.18	206.03
E20r.1n	−671.9	11.27	509.27	206.07
E20r.1r	−705.5	11.71	509.70	206.24
E20n	−728.3	12.00	510.00	206.36
E19r	−822.5	13.69	511.69	207.05
E19n	−843.4	14.06	512.06	207.20
E18r	−898.6	15.01	513.01	207.58
E18n	−989.6	16.31	514.31	208.10
E17r	−1183.4	19.73	517.73	209.49
E17n	−1237.9	20.87	518.87	209.95
E16r	−1269.3	21.60	519.60	210.25
E16n	−1522.6	26.05	524.05	212.05
E15r.2r	−1584.3	26.82	524.82	212.36
E15r.1n	−1590.5	26.90	524.90	212.40
E15r.1r	−1631.2	27.41	525.41	212.60
E15n	−1756.6	29.47	527.47	213.44
E14r	−2044.2	33.15	531.15	214.92
E14n	−2289.8	36.19	534.19	216.16
E13r	−2459.4	38.20	536.20	216.97
E13n.2n	−2664.3	40.48	538.48	217.89
E13n.1r	−2671.4	40.57	538.57	217.93
E13n.1n	−2772.0	41.87	539.87	218.46
E12r	−2933.4	43.93	541.93	219.29
E12n	−2962.8	44.35	542.35	219.46
E11r	−3266.9	49.31	547.31	221.47
E11n	−3305.1	49.99	547.99	221.75
E10r	−3382.4	51.22	549.22	222.24
E10n	−3430.4	52.43	550.43	222.74
E9r	−3559.7	55.57	553.57	224.01
E9n	−3613.9	56.88	554.88	224.54
E8r	−3773.6			226.15
E8n	−3832.4			226.75
E7r	−3868.3			227.12
E7n	−3993.9			228.41
E6r	−4028.6			228.76
E6n	−4096.5			229.46
E5r	−4134.4			229.85
E5n.2n	−4148.6			229.99
E5n.1r	−4159.6			230.10
E5n.1n	−4198.2			230.50
E4r	−4215.4			230.67
E4n	−4256.1			231.09
E3r	−4272.3			231.26
E3n	−4295.1			231.49
E2r	−4329.6			231.84
E2n	−4368.6			232.24
E1r (*partim*)	−4400.9			232.57

Notes to Table 3

Polarity chrons are numbered upward from base of section recovered in NBCP cores to CAMP volcanics in Newark basin (E1r to E24n) and from H24n (equivalent to E24n) to H27n in Hartford basin with suffix n for normal polarity and r for reverse polarity. Stratigraphic levels (arbitrarily made negative for Newark basin where scaled to the Rutgers NBCP core and positive in Hartford basin) are measured with respect to the base of the Orange Mountain Basalt (OMB) and equivalent Talcott Basalt, which is given for reference. The position of the base of each chron is also given within the nearest member as McLaughlin cycle (McL), whose base is assumed to be in more or less consistent phase with a peak of the 405 kyr orbital eccentricity cycle. The age of each polarity chron is estimated from its fractional position in the inferred constituent 405 kyr cycle (Ecc405) counted back (k) from its most recent peak (k = 1) at 0.216 Ma, i.e., Age (Ma) = $0.216 + (k - 1) * 0.405$. Ages of polarity chrons E8r to E1r (base not recovered) in lower Stockton Formation in Princeton NBCP core are estimated by extrapolation of sediment accumulation rate for RaR-1 to RaR-8. Values of the stratigraphic levels of the polarity chrons in the Hartford basin have been modified as in Table 2 but with no change in ages from Kent and Olsen (2008).

3. Integration with marine sections

Standard divisions of the Mesozoic time-scale–in this case, the Carnian, Norian, and Rhaetian stages of the Late Triassic and the Hettangian and Sinemurian epochs of the Early Jurassic–are typically delineated in marine sediments on the basis of ammonoid/ammonite or conodont biostratigraphies. With the paucity of radioisotopically dated horizons or clear Milankovitch signatures in most marine sections, magnetostratigraphy has become a powerful tool to correlate the marine sections to the Newark-Hartford APTS to develop an integrated global chronology. Marine sections in the Tethyan realm are often rich in age-diagnostic fossils and have been a favored target of magnetobiostratigraphic studies (see the comprehensive review of Triassic magnetobiostratigraphic sections in Hounslow and Muttoni (2010).

Two issues have stood out for Late Triassic time-scales: the long Carnian versus long Norian options, and the short versus long Rhaetian options. The magnetostratigraphic correlations and new age dates summarized below strongly support a relatively long Norian with a (redefined) Rhaetian of modest duration.

In the early stages of the development of what we now refer to as the Newark-Hartford APTS (Kent and Olsen, 1999; Kent et al., 1995; McIntosh et al., 1985; Witte et al., 1991), there was a reliance on palynofloral zonations for stage-level assignments (e.g., Cornet and Olsen, 1985). Moreover, some of the best available time-scales at the time made the entire Late Triassic relatively short, for example, only 21.7 Myr long, from 227.4 Ma for the base of the Carnian to 205.7 Ma for the TJB with the Carnian/Norian boundary at 220.7 Ma and the Norian/Rhaetian boundary at 209.6 Ma in Gradstein et al. (1994). It was soon recognized that the TJB and onset of volcanism in the early part of Chron E24n was actually closer to 202 Ma than 205.7 Ma. But for lack of better dating and means of correlation, the Carnian/Norian boundary, which in some time-scales was as young as 215 Ma (Webb, 1981), was placed at the New Oxford-Lockatong/Lower Passaic-Heidlersburg palynofloral zonal boundary (Fig. 1**),** more or less coincident with Chron E13 whose astrochronostratigraphic age was within the then-permissible bounds of 215 and 220 Ma. In similar fashion, the Norian/Rhaetian boundary was equated with the Lower Passaic-Heidlersburg/Upper Balls Bluff-Upper Passaic palynofloral zonal boundary at around Chron E18 (ca. 208 Ma) whereas the base of the Carnian was left as a question mark because no pre-Carnian palynofloral elements were recognized even in the lowest part of the Newark section, giving the impression that the Carnian could be 15 Myr or longer in duration.

The breakthrough came with the magnetobiostratigraphies of Silicka Brezova (Channell et al., 2003) and Pizzo Mondello (Muttoni et al.,

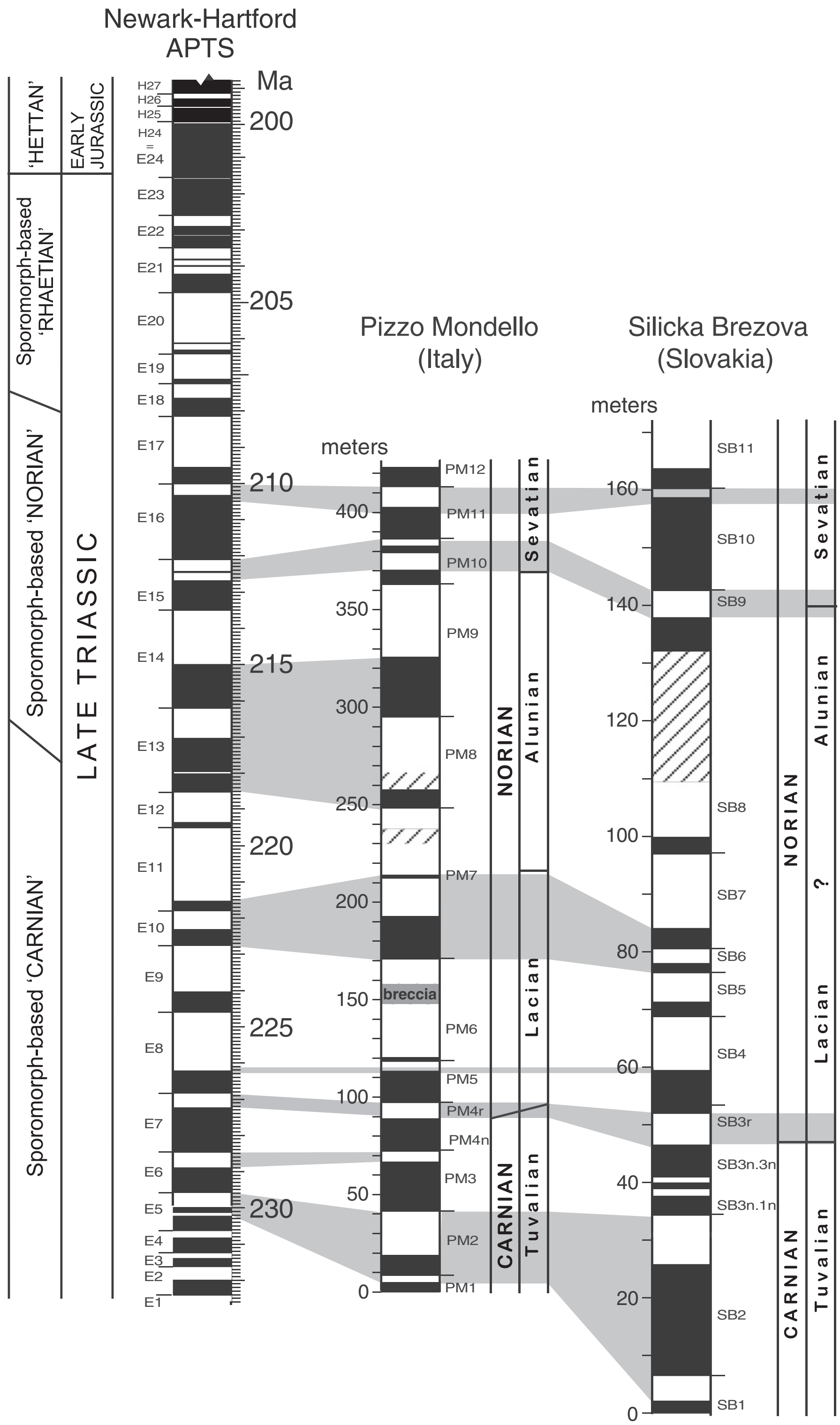


Fig. 10. Correlation of magnetobiostratigraphies from Pizzo Mondello (Muttoni et al., 2004a) and Silicka Brezova (Channell et al., 2003) with the Newark-Hartford APTS. Correlation as in (Muttoni et al., 2004a) from PM4r = SB3r (Carnian/Norian boundary interval) = E7r to PM12 = SB11 = E17. Correlation of Pizzo Mondello to the Newark-Hartford APTS below PM4r = E7r are slightly modified after (Muttoni et al., 2004a; see also Fig. 14). Substages (Tuvalian, Lacian, and Alaunian) are established using conodont biostratigraphy, updated for Pizzo Mondello by (Mazza and Rigo, 2012). Figure modified from Olsen et al. (2011).

2004a) and their correlations to the Newark APTS (Fig. 10). Both of these independent studies correlated the conodont-based Carnian/Norian boundary in these marine sections much further back in the Newark APTS, to Chron E7 with an astrochronostratigraphically estimated age of about 227 Ma. This immediately lengthened the duration of the Norian by 7–10 Myr at the expense of the Carnian and became the long Norian option. A subsequent astronomical cycle match between Pizzo Mondello and the Newark-Hartford APTS (Hüsing et al., 2011) was found to be consistent with the long-Norian magnetostratigraphic correlation scheme proposed by Muttoni et al. (2004a) (Fig. 11).

Running in parallel with debate about the length of the Norian have been questions about the duration of the Rhaetian, from approximately

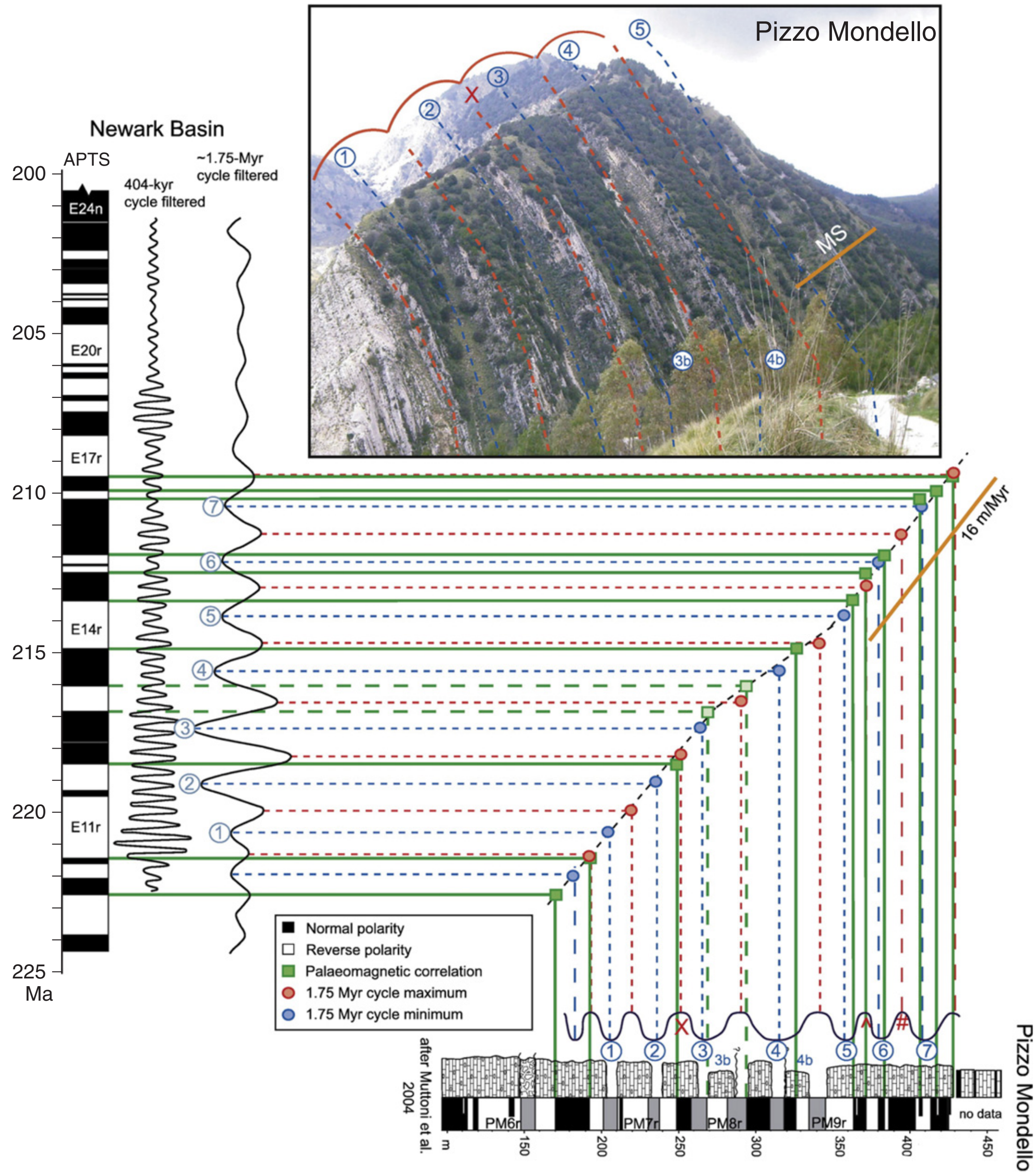


Fig. 11. Sedimentary record of climate cycles at Pizzo Mondello (outcrop section shown in photograph at top) correlated to portion of Newark-Hartford APTS. Figure edited from Hüsing et al. (2011).

2 Myr (Callegaro et al., 2012; Gallet et al., 2007) to something more like 9 Myr (Muttoni et al., 2010). This wide disparity in large part was (and continues to be) due to varying biostratigraphic criteria for recognition of the Norian/Rhaetian boundary. Correlation of the magnetobiostratigraphy of the Global Stratotype Section and Point (GSSP) candidate for the base of the Rhaetian at Steinbergkogel (Austria) to Chron E16r of the Newark APTS (Hüsing et al., 2011) suggested an age close to 209.5 Ma (Fig. 12). This was consistent with the long Rhaetian option and adopted, for example, in the Geological Time Scale 2012 (Gradstein et al., 2012) although the date is shown as 208.5 Ma in a more recent (v. 2016/12) IUGS/ISC chronostratigraphic chart (www.stratigraphy.org). Because of taxonomic issues regarding the nominate conodont species (*Misikella posthernsteini*), whose first appearance datum is supposed to coincide with the base of the Rhaetian, Maron et al. (2015) proposed using an alternative chemostratigraphic criterion (negative shift of ca. 6‰ of the $\delta^{13}C_{org}$ occurring 50 cm below the first appearance of *M. posthernsteini* sensu stricto) in the GSSP candidate at Pignola-Abriola (Italy) that can be magnetostratigraphically correlated to Chron E20r.2r at 205.7 Ma. This alternative criterion for the base of the Rhaetian, which has been proposed as the basis of the Norian/Rhaetian boundary GSSP at the Pignola-Abriola section (Bertinelli et al., 2016), is important to bear in mind for interpreting biostratigraphically-calibrated U-Pb dates discussed below.

Better defined are the marine-terrestrial correlations for the Hettangian and the Hettangian/Sinemurian boundary because the GSSP and criteria for definition have been already agreed upon (Bloos and Page, 2002). The magnetic polarity pattern is rather simple in the Hartford section: a 1.6 Myr-long normal polarity Chron H24n (= E24n) encompassing the CAMP interbedded basalts and a million years of subsequent sediment deposition is followed by three relatively short reverse polarity interval (H25r, H26r and H27r) (Kent and Olsen, 2008) (Fig. 3). An independent astronomically-calibrated magnetobiostratigraphy for Hettangian and Sinemurian marine sections at St. Audrie's Bay and East Quantoxhead in Somerset, United Kingdom (Hüsing et al., 2014) agrees remarkably well with the Hartford

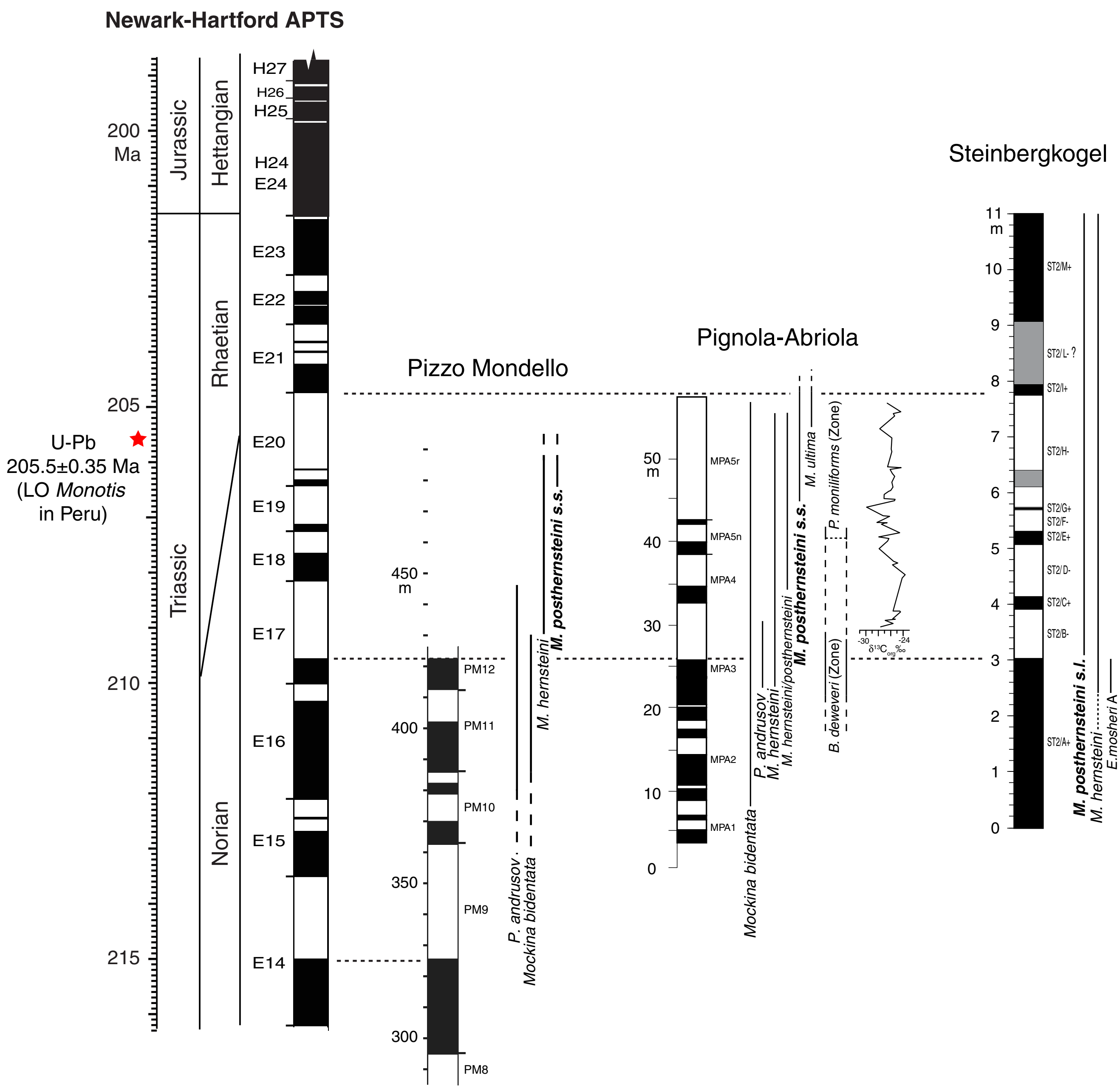


Fig. 12. Constraints on the Rhaetian from correlation of magnetobiostratigraphies at Pizzo Mondello (Muttoni et al., 2004a), Pignola-Abriola (Maron et al., 2015), and Steinbergkogel (Hüsing et al., 2011) to the Newark-Hartford APTS. Figure adapted from Maron et al. (2015).

terrestrial record at better than the 405-kyr long eccentricity level (Fig. 13). Moreover, the St. Audrie's Bay–East Quantoxhead magnetobiostratigraphic record encompasses the GSSP for the Hettangian-Sinemurian boundary (Bloos and Page, 2002), which falls close to magnetozone AQ2r, correlative to Chron H25r (Hüsing et al., 2014).

4. Comparison of Newark-Hartford APTS with other U-Pb dates

The 32.5 Myr-long Newark-Hartford APTS is fundamentally a time series of orbitally-forced climatic effects that are assumed to be both accurately recorded and recognized in the geological record. This floating astrochronology is anchored only at one level, near the younger end of the sequence by U-Pb zircon dates for the onset of CAMP volcanics just after Chron E23r. This simple scheme makes the Newark-Hartford APTS amenable to verification by independent high-precision U-Pb dates. Correlations to the Newark-Hartford APTS also make testable predictions about the relative durations of geologic stages. The challenge has been to find dated levels that can be precisely correlated to the Newark-Hartford APTS.

4.1. Carnian/Norian boundary

A U-Pb zircon date of 237.77 ± 0.14 Ma for an ash layer in the last ammonoid subzone of the Ladinian Stage (regoledanus) from the Rio Nigra section close to the Ladinian-Carnian GSSP at Stuores (Mietto et al., 2012) points to an age of nearly 237 Ma for the base of the Carnian (Fig. 14; ; see also paragraph 5.2). Three marine hosted U-Pb zircon dates of middle to early Norian age according to conodonts are 223.8 ± 0.74 and 224.47 ± 0.29 Ma from British Columbia (Daikow et al., 2011; Diakow et al., 2012), and 225 ± 3 Ma from Alaska (Gehrels et al., 1987). The two dates from British Columbia, although

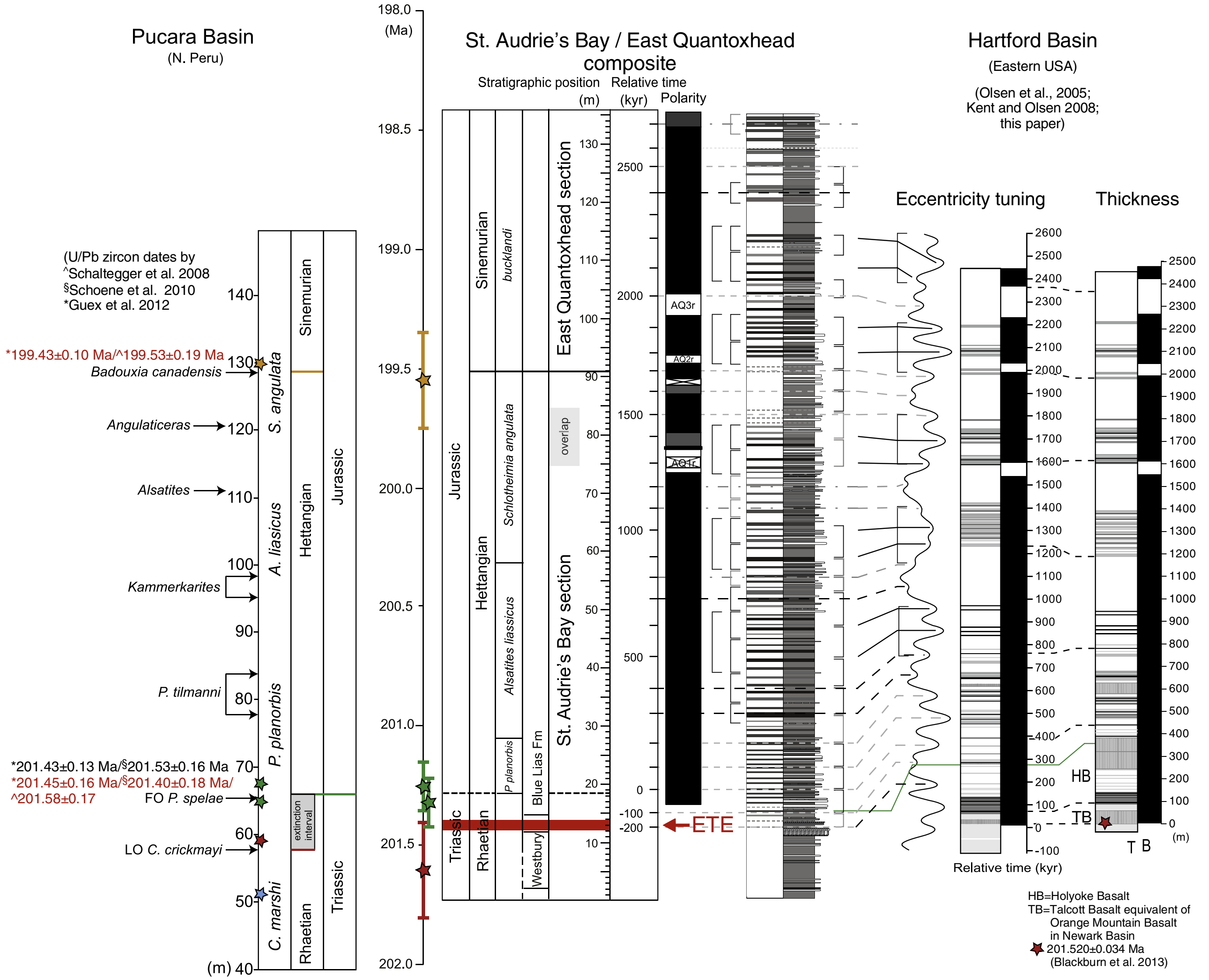


Fig. 13. Correlation of magnetobiostratigraphy and cycles of St. Audrie's Bay / East Quantoxhead composite section to the Newark-Hartford APTS (adapted from Hüsing et al., 2014). The Hartford basin columns have been updated by taking into account new data from cores BD 227A and 255 (Steinen et al., 2015), which added a net 34.07 m to the basal East Berlin and overlying section but with no change to the astrochronological ages of member boundaries (see Fig. 3 and Table 2).

yet only published in abstract, are independent and agree with the date albeit of low precision from Alaska. According to these dates, together with a late Carnian (late Tuvalian, based on conodonts) U-Pb zircon date of 230.91 ± 0.33 from marine strata of the Pignola 2 section in Italy (Furin et al., 2006), the Carnian/Norian boundary is constrained to be between 225 and 231 Ma. These data alone are compatible with only the long Norian option.

The older end of the Newark-Hartford APTS is based on extrapolation of sediment accumulation rates in the Raven Rock members of the Stockton Formation back from around 226 Ma to 233 Ma (Fig. 9). The magnetostratigraphic correlations of the Silicka Brezova (Channell et al., 2003) and Pizzo Mondello (Muttoni et al., 2004a) Tethyan marine sections to the Newark basin APTS suggest the Carnian/Norian boundary should be broadly within Chron E7r at about 227 Ma (Fig. 10). Correlation of the U-Pb zircon dated, conodont-bearing Pignola 2 section from Italy to the Newark-Hartford APTS provides independent insights on the age of the base of the Newark basin APTS (Furin et al., 2006). According to Furin et al. (2006), the biostratigraphic correlation using conodonts between Pignola 2 and Pizzo Mondello indicates that the 5-cm-thick Aglianico volcanic ash bed with a U-Pb zircon date of 230.91 ± 0.33 Ma should fall in the lowermost part of the Pizzo Mondello section and within Chron E6 of the Newark-Hartford APTS at around 229 Ma. The conodont biostratigraphy of Pignola 2 has been however recently revised (Rigo et al., 2012). So was the biostratigraphy of Pizzo Mondello (Mazza and Rigo, 2012), a revision resulting in a more expanded Carnian assemblage and a confirmation of the position of the Carnian/Norian boundary within PM4r (as approximated for example by conodonts *Metapolygnathus communisti* or *M. parvus*) correlative to Chron E7r in the Newark APTS as originally suggested by Muttoni et al. (2004a). A selection of revised Carnian conodonts from both sections, including *M. praecommunisti*, *Paragondolella noah*, *P. oertlii*, and *Carnepigondolella carpathica*, has been used to erect an informal concomitant range zone that seems to indicate that the base of Pizzo Mondello should not extend below about the 15 m level at Pignola 2 and as a consequence the dated ash bed should actually fall below the base of Pizzo Mondello. This level projected on the Newark-Hartford APTS following the Pizzo Mondello to Newark correlation of Muttoni et al. (2004a) and this study for magnetozones older than PM4r falls broadly within Chrons E5–E4 at around 231 Ma (Fig. 14). This is in substantial agreement with - and thus serves to confirm the -(extrapolated) astrochronological ages of the Newark-Hartford APTS.

4.2. Norian/Rhaetian boundary

More complicated is the interpretation of U-Pb dates for the Norian/Rhaetian boundary and the resulting duration of the Rhaetian (Fig. 15). A long (6–9 Myr) Rhaetian option was favored from the frequency of polarity reversals as well as their correlation to the Newark-Hartford APTS of a 500 m-thick composite section of Rhaetian strata in the Southern Alps (Muttoni et al., 2010). Despite the prevalence of hardgrounds (hiatuses) in the slowly accumulating Halstatt facies, detailed sampling of the nominal 10 m-thick proposed Rhaetian GSSP section at Steinbergkogel in Austria also showed a sufficiently large number of polarity reversals to favor a relatively long (of order 10 Myr) Rhaetian (Hüsing et al., 2011). In contrast, the nominally 20 m-thick section at Oyuklu in Turkey showed only a few polarity changes and this helped inspire the concept of a short (only about 2 Myr) Rhaetian (Gallet et al., 2007).

U-Pb zircon dates from volcanic ash layers within strata recording the last occurrence of large monotid bivalves in the Pucara Basin in Peru suggest an age between 205.70 ± 0.15 Ma and 205.30 ± 0.14 Ma for the Norian/Rhaetian boundary (Wotzlaw et al., 2014). This age estimate is seriously at odds with the age of around 209.5 Ma obtained by correlation of the magnetostratigraphy of the candidate Rhaetian GSSP at Steinbergkogel to the Newark-Hartford APTS (Hüsing et al., 2011). However, an alternative chemostratigraphic criterion for the base of the Rhaetian at the Pignola-Abriola candidate GSSP shows that the Norian/Rhaetian boundary can be correlated to Chron E20r.2r at 205.7 Ma (Maron et al., 2015) (Fig. 15). This age is coherent with the U-Pb zircon dates from Peru (Wotzlaw et al., 2014) and together support a modest duration of about 4 Myr for the Rhaetian. But rather than implying a long cryptic hiatus in the Newark section, whereby the Rhaetian as traditionally (yet variously) defined is supposed to be largely missing in the Newark basin section, the mutually consistent results of Wotzlaw et al. (2014) and (Maron et al., 2015) simply mean that the palynofloral zones in the Newark Supergroup, for want of better criteria at the time for correlations with marine sections (Cornet, 1993; Cornet and Olsen, 1985), do not coincide with standard geologic stages based on marine biostratigraphy.

4.3. Triassic/Jurassic and Hettangian/Sinemurian boundaries

Assessments of the ages of the Triassic/Jurassic and Hettangian/Sinemurian boundaries also come from U-Pb zircon dates on volcanic ash layers in fossiliferous marine sections of the Pucara Basin in Peru (Fig. 13). Dates that bracket the first occurrence of *Psiloceras spelae*, the nominate datum for the base of the Jurassic at the GSSP at Kujhoch, Austria (Hillebrandt et al., 2013), indicate an age of 201.31 ± 0.18 Ma for the TJB (Schoene et al., 2010), with supportive U-Pb zircon dates in the context of a detailed ammonite biostratigraphy (Guex et al., 2012). The age for the TJB from Peru is in good agreement with or perhaps slightly younger than the oldest CAMP U-Pb dates, such as the Palisade Sill/Orange Mountain Basalt (201.520 ± 0.034 Ma) and North Mountain Basalt (201.566 ± 0.031 Ma) (Blackburn et al., 2013). This would imply that the TJB post-dated the inception of CAMP activity. On the other hand, the ETE seems to just precede or is nearly coincident with the onset CAMP activity in North America and Morocco (Blackburn et al., 2013). Based on projection from the St. Audrie's Bay section (first occurrence of *Psiloceras planorbi*s - assumed to be close to that of *P. spelae*), the TJB should be at 201.42 ± 0.02 in the Newark-Hartford APTS (Sha et al., 2015). This is within error of the 201.36 ± 0.14 Ma average of the bracketing dates from Peru (Wotzlaw et al., 2014; Sha et al., 2015).

A U-Pb zircon date of 199.53 ± 0.19 Ma from a tuff within the *Badouxia canadensis* beds that should closely approximate the Hettangian/Sinemurian boundary indicates, in concert with the U-Pb zircon dates also from Peru for the TJB, that the Hettangian is only about 2 Myr long (Schaltegger et al., 2008). This is in excellent agreement with the duration of the Hettangian determined by astrochronology in the St. Audrie's Bay–East Quantoxhead composite section (Fig. 13) and is fully consistent with the Newark-Hartford APTS (Hüsing et al., 2014).

5. Extension of the Newark-Hartford APTS into younger and older strata

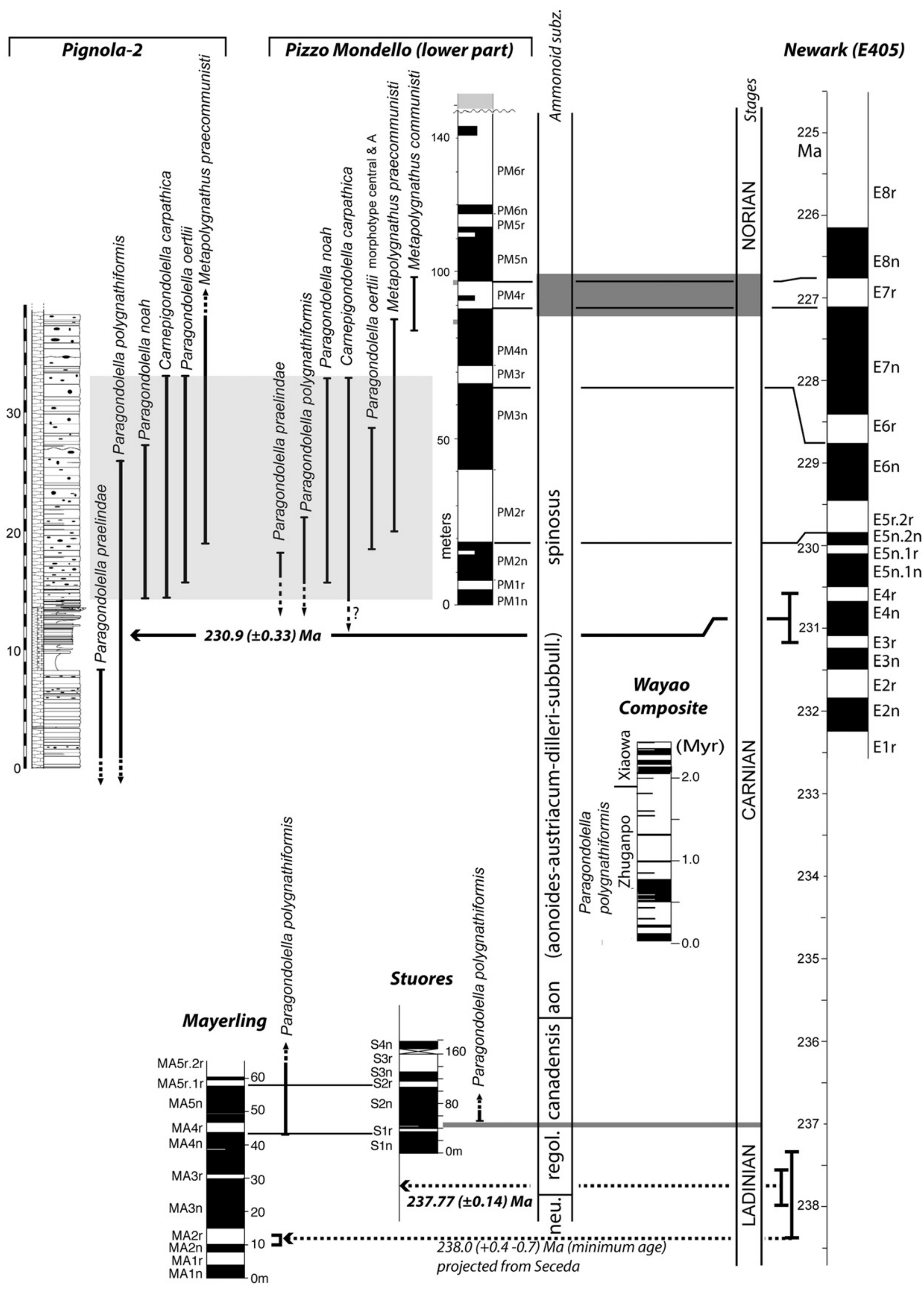
5.1. Sinemurian and younger

There is another 2000 m of section above the sampled interval of the Portland Formation in the Hartford basin. Cyclicity, however, is less apparent and the lithology becomes dominated by more eolian facies (Hubert et al., 1992), making it difficult to map distinctive units, establish accurate stratigraphic placement of sampling sites, as well as to find suitable lithologies there to extend the magnetostratigraphy upward into and above Chron H27n. The marine sections at St. Audrie's Bay–East Quantoxhead (Britain) offer better prospects for extending the polarity sequence into Sinemurian and younger strata as well as for a bridge to Sinemurian and Pliensbachian (to about 183 Ma) magnetostratigraphies from the Umbrian Apennines (Fonte Avellana and Cingoli; Channell et al., 1984) and the southern Italian Alps (Colle di Sogno; Channell et al., 2010), even though radiometrically dated levels or cycle stratigraphic age controls are apparently not yet available. In any case, there is still a sizable age gap to link these magnetostratigraphic records to the Late Jurassic-Early Cretaceous oceanic anomaly M-sequence, which is relatively secure back to Chron M29 or about 155 Ma (Malinverno et al., 2012; Tominaga and Sager, 2010) and verified by magnetostratigraphy (Channell et al., 1995, 2010; Speranza et al., 2005). However, even the basic polarity sequence is far less certain in the Middle Jurassic where the oceanic record is characterized by small-scale magnetic anomalies of continued decreasing amplitude (Cande et al., 1978) that are now numbered back to M44 with an extrapolated age of about 170 Ma (Sager et al., 1998; Tominaga et al., 2008) but of uncertain origin (Gee and Kent, 2007; McElhinny and Larson, 2003). The available Middle Jurassic magnetostratigraphic records tend to be very noisy with many apparent flips in polarity, making correlations uncertain (e.g., Steiner et al., 1987).

5.2. Carnian and older

Stratigraphic sections with age control older than the base of the Newark-Hartford APTS (extrapolated age of about 232.5 Ma) within the Carnian are Mayerling in Austria (Gallet et al., 1998) and the GSSP for the Ladinian/Carnian boundary at Stuores in Italy placed at the regoledanus/canadensis ammonoid subzone boundary (Mietto et al., 2012). The Stuores GSSP has been tentatively correlated to the nearby Rio Nigra section that provided a U-Pb zircon date of 237.77 ± 0.14 Ma for an ash layer located in the regoledanus subzone just above the neumayri/regoledanus subzone boundary, approximately one ammonoid subzone below the Ladinian/Carnian (regoledanus/canadensis) boundary estimated at about 237 Ma (Mietto et al. (2012) and references therein) (Fig. 14). These and several other Middle–early Late Triassic sections including the Seceda core (Muttoni et al., 2004b) were used to erect a radiometrically constrained (U-Pb zircon dated) magnetobiostratigraphy extending from the late Anisian into the Carnian (see Fig. 5 in Hounslow and Muttoni (2010) and references therein), with the Anisian/Ladinian boundary presently estimated at 242 Ma (Mundil et al., 2010). A U-Pb zircon date of 238.0 (+0.4/−0.7) Ma from an ash bed within the Ladinian archelaus ammonoid zone (whose upper part corresponds to the neumayri subzone) at Seceda, considered a minimum age because potentially affected by Pb loss (Mundil et al., 2010), when projected onto Mayerling using the correlations scheme of Hounslow and Muttoni (2010) provides additional support for a Ladinian/Carnian boundary age at close to 237 Ma (Mietto et al., 2012).

The gap between Stuores-Mayerling and the Carnian-Norian record at Pizzo Mondello (see also above) is partially bridged by a cycle-calibrated magnetostratigraphy of Carnian carbonates from South China (Minzoni et al., 2015; Zhang et al., 2015) (Fig. 14). The 210-m-thick Wayao composite magnetostratigraphy encompassing the Zhuganpo and Xiaowa formations has been astrochronologically calibrated using long and short eccentricity resulting in a floating magnetochronology spanning about 2.4 Myr (Zhang et al., 2015). It is not obvious how to insert this floating chronology into our Carnian correlation framework because the biostratigraphic age constraints of the



investigated units are very sparse. The Zhuganpo Formation is considered Carnian in age based essentially on the presence of conodont *Metapolygnathus* (*Paragondolella*) *polygnathiformis* (Zhang et al., 2015), which first appears at Stuores close to the base Carnian level and is last present at Pignola 2 and Pizzo Mondello in levels below the base of the Norian. According to these data, we tentatively consider the 1.3 Myr-long reverse polarity interval encompassing the upper Zhuganpo and the lower Xiaowa formations (Zhang et al., 2015) as a potential extension within the Carnian of Chron E1r.

6. The 'missing' Rhaetian

A persistent refrain in the literature is that most of the Rhaetian is missing in the Newark Supergroup. For example, an unconformity is placed by Kozur and Weems (2010) within a few decimeters to decameters below the CAMP volcanics practically wherever they occur (Culpeper, Gettysburg, Newark, Hartford and Fundy basins, their Figs. 9 & 10) on the basis of conchostracan biostratigraphy. Tanner and Lucas (2015) assert that correlation of "pollen and conchostracan zones between the Newark Supergroup and the Germanic Triassic, indicates that most of Rhaetian and a portion of late Norian time is not represented by sediment in the Newark basin and elsewhere in the Newark Supergroup." Van Veen (1995) argues that the reported palynofloral change just below the North Mountain Basalt and shortly before the end-Triassic extinction in the Fundy basin (Fowell and Olsen, 1993) actually took place 2–3 Myr earlier based on evidence from Western Europe, a viewpoint that was shared at the time by Kuerschner et al. (2007) from work on the Tiefengraben section in Austria. Finally, Gallet et al. (2007) suggest that the Rhaetian is at least partly missing in the Newark basin based on magnetostratigraphic correlations with a condensed marine section in Turkey (Oyuklu).

The purported hiatus in the Newark section is often intermingled with the so-called short-Rhaetian option (Ogg, 2012). When the temporal extent of the hiatus according to various workers is shown explicitly, as by Wotzlaw et al. (2014), it is placed below the CAMP lavas, squeezed into the lowest part of Chron E24n and just above Chron E23r (Fig. 15). The hiatus effectively is made to coincide with the ETE. In fact, Tanner and Lucas (2015) reassign Chron E23r from the late Rhaetian to the late Norian on the basis of the conchostracan *Shipingia olseni* immediately below and with the late Norian palynomorphs (e.g. *Patinasporites densus*) above this magnetozone. However, the recorded presence of the very short (nominally 10 kyr) Chron E23r immediately below CAMP basalts in three entirely separate basins, namely Newark (Kent and Olsen, 1999; Kent et al., 1995; Olsen et al., 1996a), Fundy (Deenen et al., 2011) and Argana (Deenen et al., 2010) (see Fig. 8), makes such a major regional unconformity untenable. This is because the unconformity would require essentially identical intervals of nondeposition and/or erosion in all of these basins, so that somehow deposition halted immediately after or else each section was eroded to variable depths exactly down to just above Chron E23r, the shortest identified chron in the Newark-Hartford APTS. But not to be deterred, Tanner and Lucas (2015) do not regard such an argument as precluding a regional unconformity and speculate that a late Rhaetian episode of uplift along the entire rift axis might somehow have caused such precisely timed nondeposition or erosion of the exposed strata in all of the rift valleys prior to the flood basalt eruptions.

Specific evidence contradicting the notion that Rhaetian time is somehow missing in the Newark Supergroup comes from the magnetobiostratigraphy of the St. Audrie's Bay section in Britain (Briden and Daniels, 1999; Hounslow et al., 2004) (Fig. 16). The St. Audrie's Bay measured section is about 125 m thick and comprised of the Mercia Mudstone, Penarth and Lias Groups. The exposed Mercia Mudstone consists of around 100 m of reddish to grey dolomitic mudstones, siltstones and evaporites of floodplain and playa lake origin. The overlying Penarth Group, all of 13 m-thick, is divided into the Westbury Formation and overlying Lilstock Formation, which is further divided into the Cotham and overlying Langport members. The Westbury Formation and up into the lowermost Cotham Member of the overlying Lilstock Formation is reported to contain *Rhaetavicula contorta* (Golebiowski, 1990; Hounslow et al., 2004), a fossil bivalve that is known from the shallow-water facies of the Kössen Formation of the Rhaetian type area (Golebiowski, 1990; Kozur, 2003). Just above in the middle of the Cotham Member there is a prominent horizon of distinctly disturbed sediment capped by what could be taken as a bedding exposure surface with deep desiccation cracks and local evidence of erosion (Hesselbo et al., 2002; Hounslow et al., 2004; Mayall, 1983). The Cotham and the overlying Langport members do contain some marine fossils, notably conodonts (e.g., *Chirodella verecunda*) that occur just below the top of the Langport Member at Lilstock (Swift and Martill, 1999). These conodonts occur in association with *Misikella conformis* and *M. posthernsteini* in the Langport Member and the immediately overlying pre-*planorbis* beds, respectively, and are indicators of a Rhaetian age (Hounslow et al., 2004). The Lias Group consists of limestones and shales with the diagnostic Early Jurassic ammonoid (*P. planorbis*) just above the pre-*planorbis* beds (PPB in Fig. 16) of the Blue Lias Formation at the top of the magnetostratigraphic section.

Within these broad biostratigraphic constraints, for example, that the ETE and TJB must occur somewhere between the top of the Lilstock Member and the base of the *P. planorbis* subzone of the Blue Lias Formation (i.e., presumably within the pre-*planorbis* beds), a satisfactory correlation can be made of the St. Audrie's magnetostratigraphy to the Newark-Hartford APTS, especially for the Mercia Mudstone Group initially studied by Briden and Daniels (1999) and sampled further by Hounslow et al. (2004). According to these studies, the Mercia magnetostratigraphy can be convincingly correlated from Chron E14r at the exposed base of the section up into Chron E20n at the top of the unit, with the 30 m-thick interval of predominantly reverse polarity encompassing the Blue Anchor Formation diagnostically corresponding to predominantly reverse polarity Chrons E19 and E20 (Fig. 16). The sediment accumulation rate for the Mercia Mudstone would be about 10 m/Myr (100 m from 214.5–204.5 Ma). Correlation of the Penarth magnetostratigraphy is much more ambiguous but from the foregoing should correspond to most of Chrons E21 to E23. This would imply a low net sediment accumulation rate of only around 4 m/Myr (13 m of section from 204.5–201.5 Ma). Characterizing a polarity pattern in such a low sedimentation rate section is difficult in the face of sampling limitations and smearing from processes like bioturbation and haloturbation. Uncertainties in correlating the resulting polarity sequence are compounded if

Fig. 14. Correlation framework in the lower part of the Newark-Hartford APTS. The Pignola 2 section with a U-Pb zircon date of 230.91 ± 0.33 Ma (Furin et al., 2006), as indicated, and updated conodont biostratigraphy (Rigo et al., 2012) is correlated to the Pizzo Mondello magnetostratigraphy (Muttoni et al., 2004a) and updated conodont biostratigraphy (Mazza and Rigo, 2012). The Pizzo Mondello-Newark correlation is of Muttoni et al. (2004a) down to PM4r = E7r and slightly different for magnetozones below PM4r whereby PM3r is here correlated to E6r and PM2r to E5r.2r such that the Pignola 2 radiometric age (230.91 ± 0.33 Ma) is projected to fall within E4n at close to 231 Ma. The Stuores GSSP for the base of the Carnian (Mietto et al., 2012) is correlated to the coeval Mayerling section (Gallet et al., 1998) using the equivalences MA4r = S1r and MA5r.1r = S2r, a solution slightly different from Broglio Loriga et al. (1999) and Hounslow and Muttoni (2010) but that is here preferred because it brings in close correlation the first appearance at the base of the Carnian of *Paragondolella polyghathiformis*, which was only recently found at Stuores (Mietto et al., 2012). Also reported are a U-Pb zircon date of 237.77 ± 0.14 Ma for an ash layer in the late Ladinian regoledanus subzone from the Rio Nigra section close to Stuores and a U-Pb zircon (minimum) age of 238.0 +0.4 −0.7 Ma from an ash bed within the Ladinian archelaus ammonoid zone (whose upper part corresponds to the neumayri subzone) at Seceda (Mundil et al., 2010). The 2.4 Myr floating magnetoastrochronology encompassing the Carnian Zhuganpo and Xiaowa formations of China (Zhang et al., 2015) is considered to partly straddle the interval between Stuores and Pizzo Mondello sections. Ammonoid subzones: neu. = neumayiri (corresponding to the upper part of the archelaus zone) and regol. (=regoledanus); subzones in parenthesis are reported for the sake of completeness but have actually not been found in any of the considered sections.

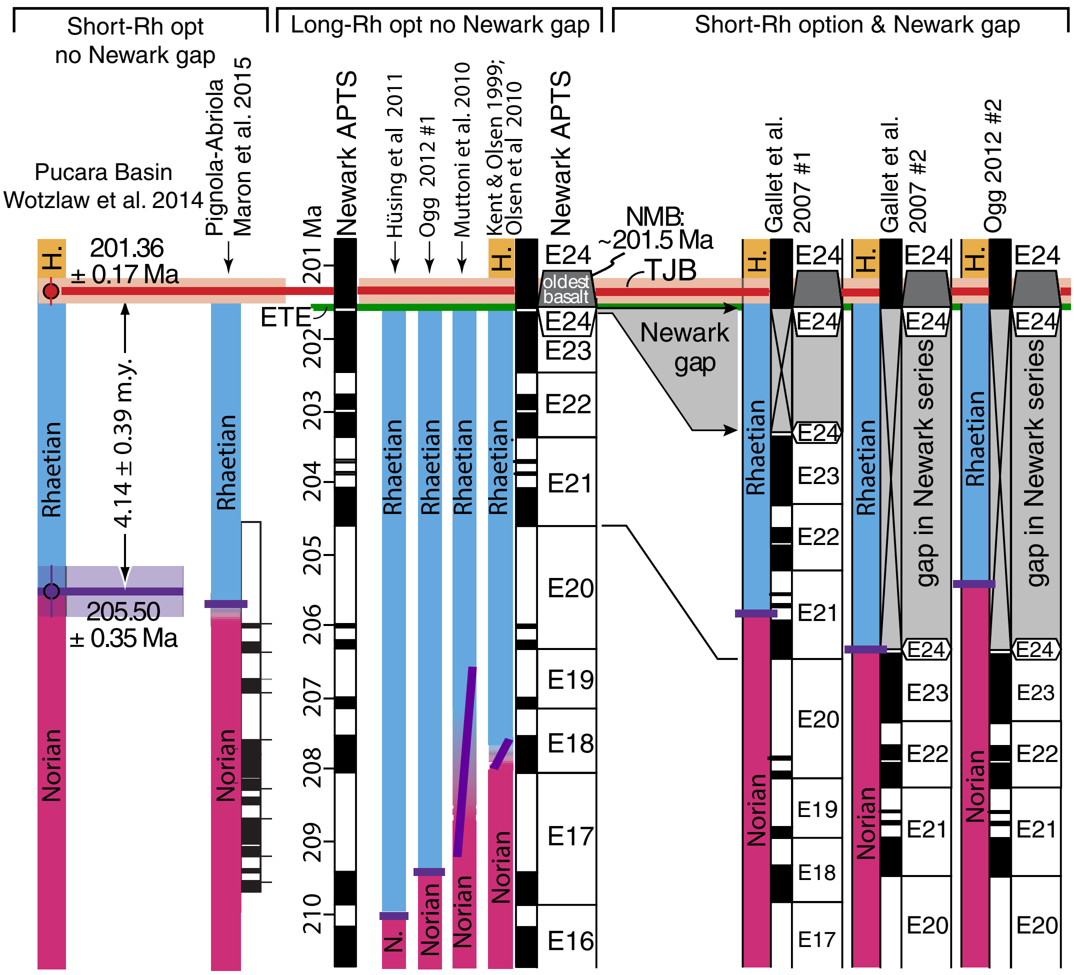


Fig. 15. Comparison of various estimates for the Norian/Rhaetian boundary and the duration of the Rhaetian (modified from Wotzlaw et al. (2014) with the addition of the entry for Maron et al. (2015) from the Pignola-Abriola section). The so-called Long-Rhaetian options were based on either correlations using sporomorphs in the Newark basin (Kent and Olsen, 1999; Olsen et al., 2011) or correlations to Newark polarity chrons (essentially the same chronology within about 0.5 Myr as the updated Newark-Hartford APTS here) using biostratigraphic criteria for the Norian/Rhaetian boundary associated with *M. posthernsteini* s.l. (Hüsing et al., 2011; Muttoni et al., 2010; Ogg, 2012), whereas, as pointed out by Maron et al. (2015), a correlation using *M. posthernsteini* s.s in the Pignola-Abriola section in the equivalent of Chron E20r places the boundary close in numerical age to the mean U-Pb zircon date (205.50 ± 0.35 Ma) for the last occurrence of the bivalve *Monotis* in Peru (Wotzlaw et al., 2014; see also Fig. 12). Note that the short-Rhaetian options invariably insert a hiatus up to 5 Myr long in the Newark basin for which we say there is no direct evidence, as discussed in the text. The Pignola-Abriola section has been proposed as the Norian/ Rhaetian GSSP (Bertinelli et al., 2016). See also Golding et al. (2016) for U-Pb zircon geochronology on Rhaetian sections in British Columbia.

the Penarth is also condensed because of hiatuses, such as the mudcrack level within the Cotham Member suggests are present. Importantly, none of the reverse polarity magnetozones in the Westbury Formation are likely to represent Chron E23r, which is only about 10 kyr long according to cycle stratigraphy and thus highly unlikely to be recovered by spot sampling of a section that accumulated at only 0.4 cm/kyr. A similar argument applies to reverse magnetozone SA5r in the lowest Blue Lias Formation, whose origin and significance are enigmatic. Nonetheless, the overall magnetostratigraphic correlation of the St. Audrie's Bay section to the Newark-Hartford APTS is sufficient to establish that the interval in the upper Passaic Formation from Chrons E21 to E23 is most probably Rhaetian in age, corresponding to the reported range of *R. contorta* in the Penarth Group.

Lastly, from a purely operative standpoint, if 3 Myr or more of Late Triassic time were not recorded in the Newark basin as supposed by some authors, this would have comparable consequences in the timing of events based on anchoring the astrochronology below the alleged unconformity to the U-Pb dated CAMP lavas above it. There is no evidence of such offsets. In fact, the excellent agreement of the U-Pb zircon date of 230.91 ± 0.33 Ma in marine strata of late Carnian age (Furin et al., 2006) correlated to Chrons E4 with a Newark-Hartford APTS age in the range of 230.5–231 Ma would preclude a significant hiatus, as would the agreement with available U-Pb zircon dates for constraints on the ages for the Carnian/Norian and Norian/Rhaetian boundaries correlated to the Newark-Hartford APTS. More generally, the 70 Myr-long Triassic–Jurassic Inuyama-ATS chert sequence in Japan shows a 1.8 Myr supercycle (Ikeda and Tada, 2014) suggesting that the Newark-Hartford sequence, where the 1.8 Myr supercycle was first found (Olsen and Kent, 1999), is most probably complete.

7. pCO_2 estimates from paleosols and the $\delta^{13}C$ record from marine carbonates

Global climate change is always difficult to gauge. The Triassic practically stands alone as a period in the Phanerozoic with no evidence of persistent polar ice sheets (Frakes and Francis, 1988) and corresponds to an oscillation towards relatively warm tropical sea surface temperatures based on oxygen isotopes in calcite and aragonite shells (Veizer et al., 2000). Varying concentrations of atmospheric pCO_2 would be expected to exert a global climate signal and influence sea-surface temperatures but the role of CO_2 in climate models (Berner, 2006) is highly dependent on very few and often highly scattered estimates especially for pre-Cenozoic time (Royer, 2006).

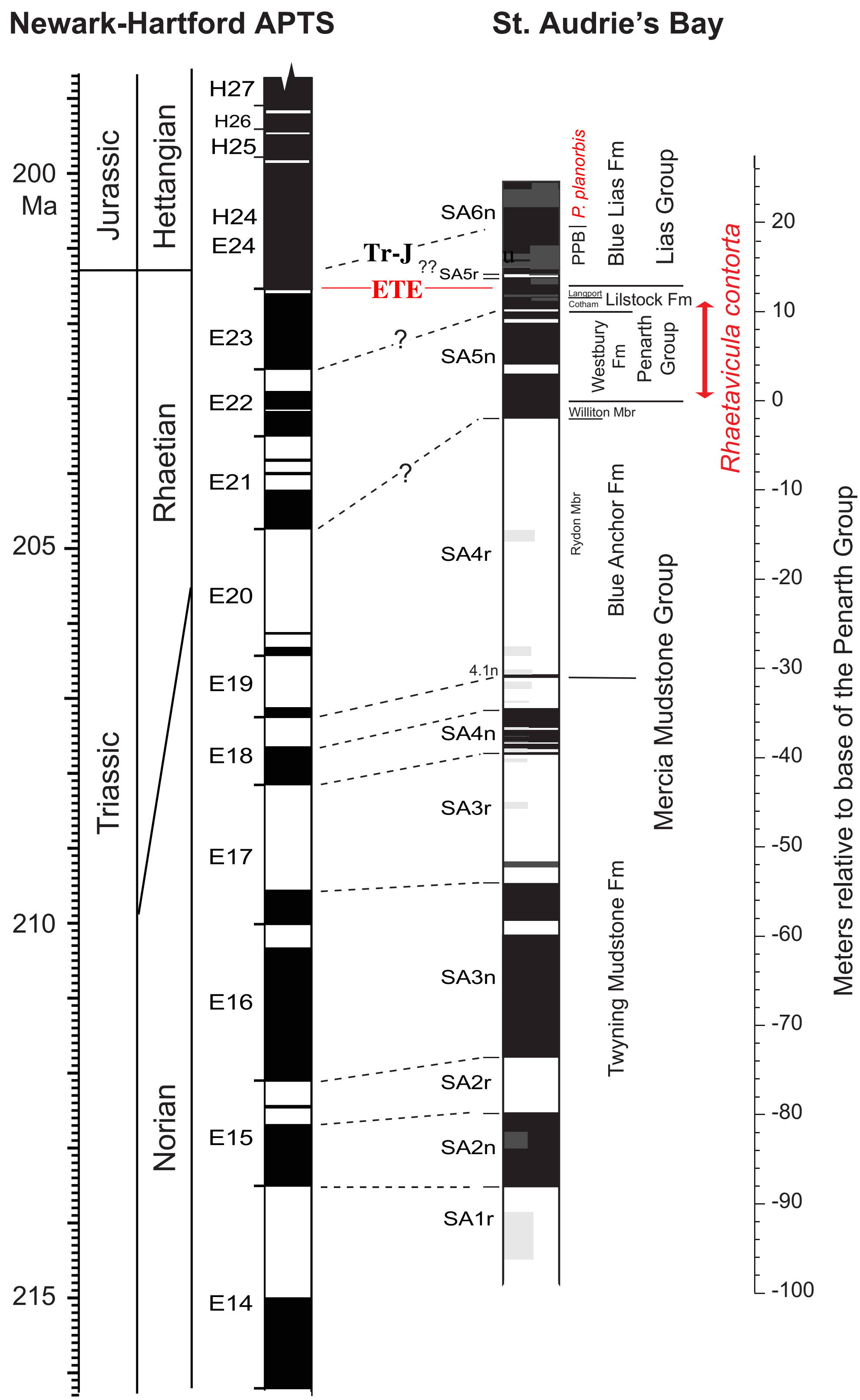


Fig. 16. Magnetobiostratigraphy of the St. Audrie's Bay section (Hounslow et al., 2004) compared to the Newark-Hartford APTS. The bivalve *Rhaetavicula contorta*, which was used to establish the Rhaetian (Kössen Beds; (Golebiowski, 1990; Kozur, 2003), occurs in the Westbury Formation (Hounslow et al., 2004) and up into the lower Cotham Member of the Lilstock Formation (Golebiowski, 1990). The magnetostratigraphic correlation of St Audrie's to Newark-Hartford APTS, either as proposed by Hounslow et al. (2004) or as suggested here, implies that Rhaetian time is represented in Newark down to around Chron E21 and perhaps even to earlier strata if the older range of *R. contorta* is shortened by significant hiatus(es) at around the base of the Penarth Group.

Paleosol carbonates can provide an excellent proxy of pCO_2 (Cerling, 1999). Paleosols are found throughout much of the Newark-Hartford cyclical facies, allowing tests of repeatability by correlation of specific Van Houten cycles within a basin and a detailed chronology by direct registry to the Newark-Hartford APTS. Estimates of pCO_2 were obtained from carbon isotope analyses of paleosol carbonates in sedimentary units interbedded with CAMP lavas to determine their outgassing potential (Schaller et al., 2011a, 2011b, 2012) and in survey mode over the entire Newark-Hartford record to establish long-term trends (Schaller et al., 2015) (Fig. 17). The results show increases in pCO_2 values by roughly a factor of two immediately after each CAMP lava unit and their subsequent decay over each ensuing 100–200 kyr interval based on the astrochronology. This provides experimental confirmation for models of the impact of a large igneous province on CO_2 outgassing and consumption by chemical weathering (Dessert et al., 2001). The long-term trends show high and steady pCO_2 values in the early to middle Norian (230–215 Ma), a dip to a low at 212 Ma and a rise back to earlier steady values by 210 Ma, followed by a steady decrease to just before CAMP eruptions. The overall pattern of atmospheric pCO_2 concentrations broadly agrees with model results based on increasing CO_2 consumption by weathering as more continental area was exposed to warm and humid equatorial climate during northward drift of Pangea (Goddéris et al., 2008; Schaller et al., 2015) but disagrees with the GEOCARBSULF carbon cycling model, which predicts steady pCO_2 values over the Late Triassic (Berner, 2006).

Variations in $\delta^{13}C$ of marine carbonates basically reflect the relative burial rate of inorganic carbon and fractionated organic carbon. The relative proportions are ultimately controlled by tectonics via nutrient supply through erosion and weathering, which can variably sustain productivity by formation of burial loci in sedimentary basins or by injection into the global carbon pool of light isotopic sources of crustal (e.g., clathrates), mantle (volcanic) or even extraterrestrial (cometary) sources. The $\delta^{13}C$ values of marine carbonates are thought to reflect global ocean isotopic compositions and can thus be useful for correlation. A suite of $\delta^{13}C$ measurements on Late (and Middle) Triassic carbonates including Pizzo Mondello, Brumano and Italcementi (Italy) that are correlated via magnetobiostratigraphy to the Newark-Hartford APTS provides a new age-calibrated target curve (Muttoni et al., 2014) (Fig. 17). The target curve shows $\delta^{13}C$ values of 2–3‰ in the Carnian punctuated by two negative excursions at around 233.5 Ma and 230 Ma, followed by a small positive excursion at 227 Ma across the Carnian/Norian boundary. The Norian record, mainly from Pizzo Mondello, is characterized by a long-term decreasing trend from average values of 2–3‰ to 1–2‰. This trend is punctuated by an additional negative excursion at around 217–213 Ma in the middle Norian. Finally, after a late Norian–early Rhaetian data gap, $\delta^{13}C$ values in the Rhaetian are on average heavier (up to 3–4‰) but display an oscillatory behavior with large and rapid increases and decreases as observed in data from Brumano and Italcementi (Muttoni et al., 2014). Aside from the uncertain origin of these oscillations, this study provides a dated $\delta^{13}C$ curve that can be useful for stratigraphic correlations and to explore possible linkages between pCO_2 and ocean carbonate $\delta^{13}C$ (Berner, 2006).

8. Correlations of continental sediments and paleogeography

Magnetic polarity and cycle stratigraphies have been especially useful tools for correlations and timing of continental successions, which typically have scant fossils of rather endemic fauna and flora. Besides the studies of the Newark and Hartford basins already described in the construction of the APTS, magnetic polarity and cycle stratigraphies have been developed from many other rift basins and allow close correlation of events. These localities include the Dan River basin of North Carolina and Virgina (Kent and Olsen, 1997; Olsen et al., 2015), the Taylorsville basin of Virginia (LeTourneau, 2003), the Jacksonwald Syncline of the Newark basin in Pennsylvania (Olsen et al., 2002a, 2002b), the Fundy basin of Nova Scotia (Deenen et al., 2011; Kent and Olsen, 2000), the Argana basin of Morocco (Deenen et al., 2010), Jameson Land in East Greenland (Clemmensen et al., 1998; Kent and Clemmensen, 1996), the Chinle Formation of the Colorado Plateau (Donohoo-Hurley et al., 2010; Molina-Garza et al., 2003; Molina-Garza et al., 1991; Steiner and Lucas, 2000; Zeigler and Geissman, 2011; Zeigler et al., 2008), the Los Colorados Formation in the Ischigualasto-Villa Union basin in Argentina (Kent et al., 2014), the Mercia Mudstone at St. Audrie's Bay in Britain (Briden and Daniels, 1999; Hounslow et al., 2004), and the long-neglected Keuper in the Germanic basin waiting to be (re)done after being the subject of a very early application of partial demagnetization techniques on the equivalent rock unit in Britain (Creer, 1959).

Paleolatitudes are typically obtained in a comprehensive magnetostratigraphic study. This information can be used to differentiate effects of global climate change (e.g., due to changes in greenhouse gases; Schaller et al., 2015) from more localized changes in environment as Pangea drifted northward across latitudinal climate belts (Kent and Tauxe, 2005). Paleogeographic reconstructions that use latitudinal data corrected for sedimentary inclination shallowing (Kent and Tauxe, 2005) are shown in (Fig. 18). It can be seen that by the end of the Triassic, the Newark basin had drifted to around 20°N within the tropical arid belt (e.g., eolian deposits in the Pomperaug basin in Connecticut (LeTourneau and Huber, 2006) and especially in the Fundy basin in Nova Scotia (Hubert and Mertz, 1980)) whereas some of the European basins had already progressed to 40° N and even higher latitudes in the temperate humid belt (e.g., fossil megafloras in Jameson Land, East Greenland and Scania, southern Sweden: McElwain et al., 1999, 2007). Kuerschner et al. (2007) had dismissed the possibility raised by Kent and Olsen (2000) that floral provincialism could account for the diachronous extinction of spore taxa. However, it seems unlikely that these disparate climatic settings in eastern North America and Greater Europe would have had many floral assemblages in common. Instead, the fossil assemblages in the different land areas may indeed be diachronous on account of the northward drift of Pangea across climate belts and provide more of an eco-stratigraphy rather than a chronostratigraphic correlation at this spatiotemporal scale.

9. Conclusions

- The Newark-Hartford APTS is paced by sixty-six (66) McLaughlin cycles, each representing climate forcing by the long astronomical eccentricity variation with the stable 405 kyr period (Laskar et al., 2011), from 199.5 to 225.8 Ma and encompassing fifty-one (51) magnetic polarity intervals (Chrons E8n to H27n). Extrapolation of sediment accumulation rates in the basal fluvial sediments (about 100 m/Myr) extends the sequence back an additional fifteen (15) polarity intervals (Chrons E1r to E7r) to 232.7 Ma. The lengths of the 66 polarity intervals vary from 0.011 to 1.63 Myr with a mean duration of 0.53 Myr.
- The Newark-Hartford APTS is anchored to high-precision U-Pb zircon dating of the oldest CAMP basalts that provide an estimated age of 201.5 Ma for the base of the stratigraphically superjacent Washington Valley Member (the base of McLaughlin cycle 61) and 201.6 Ma using cycle stratigraphy for the onset of the immediately subjacent Chron E23r.
- The base of each McLaughlin cycle is picked at the most prominent dark shale to correspond to a precession maximum close to the peak of the modulating 405 kyr eccentricity variation. The estimated age of 201.5 Ma for the base of McLaughlin cycle 61 used for calibration is remarkably consistent with the calculated phase of the 498th long eccentricity cycle counting back using a period of 405 kyr from the most recent peak at 0.216 Ma. Accordingly, we adopt a nomenclature (Ecc405:k) to unambiguously assign ages from the astrochronostratigraphy:

  $$\text{Age (Ma)} = (k-1) * 0.405 + 0.216$$

  where k is the number of 405 kyr cycles (and fraction thereof) with respect to the peak at 0.216 Ma. For example, $k = 498$ for the base of the

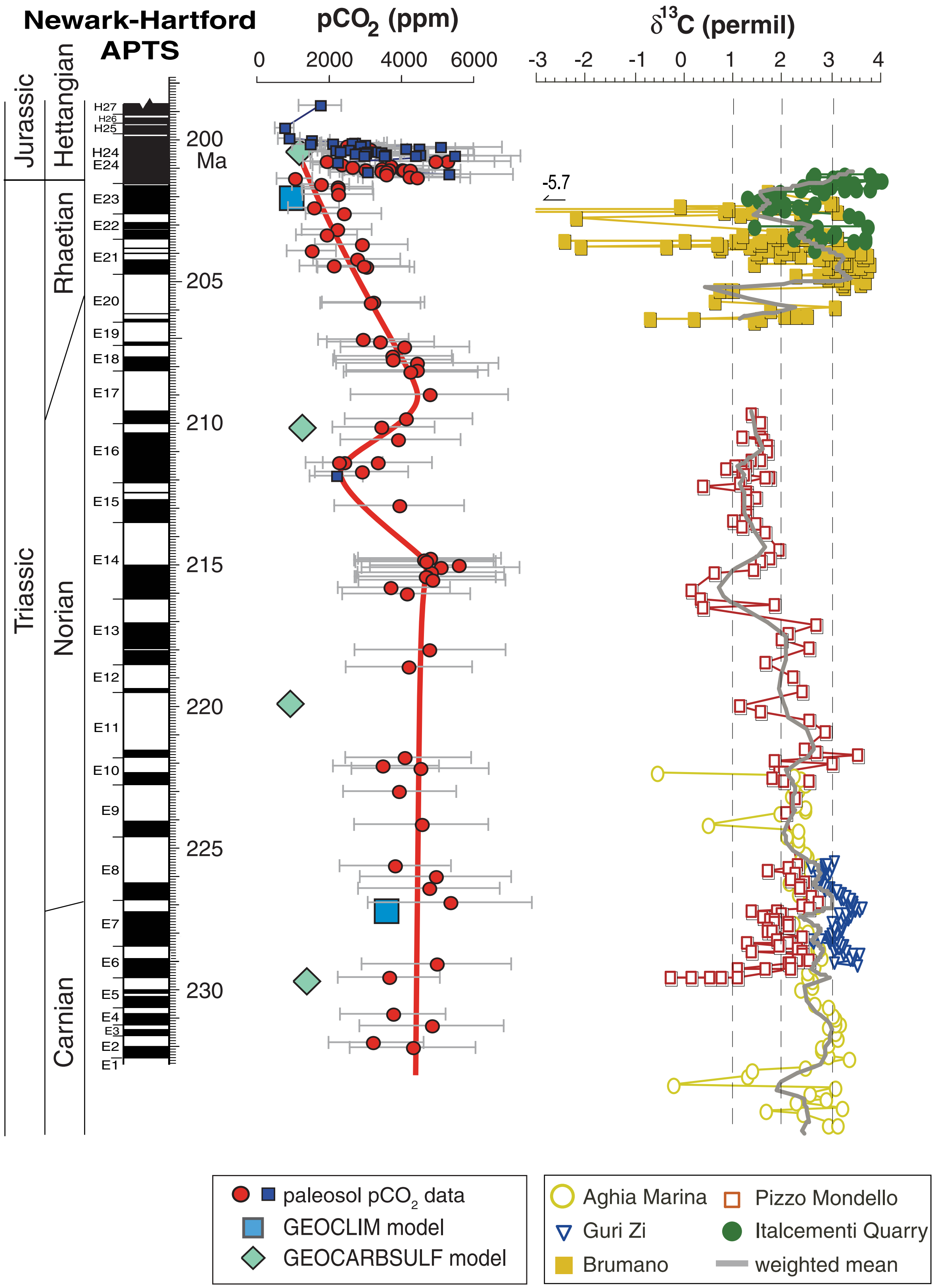


Fig. 17. Estimates of atmospheric pCO_2 from paleosol carbonates in the Newark and Hartford basins (red circles and small blue squares: Schaller et al., 2011a, 2012, 2015) and a composite carbonate $\delta^{13}C$ record from Late Triassic (Carnian–Norian–Rhaetian) marine limestones in the Tethyan realm (Muttoni et al., 2014) correlated to the Newark-Hartford APTS; see also Galli et al. (2007) and Muttoni et al. (2010) for the Triassic -Jurassic boundary interval. The pCO_2 curve is compared to results from the GEOCARBSULF model that shows little change over the Late Triassic (green diamonds; Berner, 2006) and the GEOCLIM model that shows a large decrease over 30 Myr of the Late Triassic (large blue squares; Goddéris et al., 2008), which is more consistent with the measured pCO_2 values.

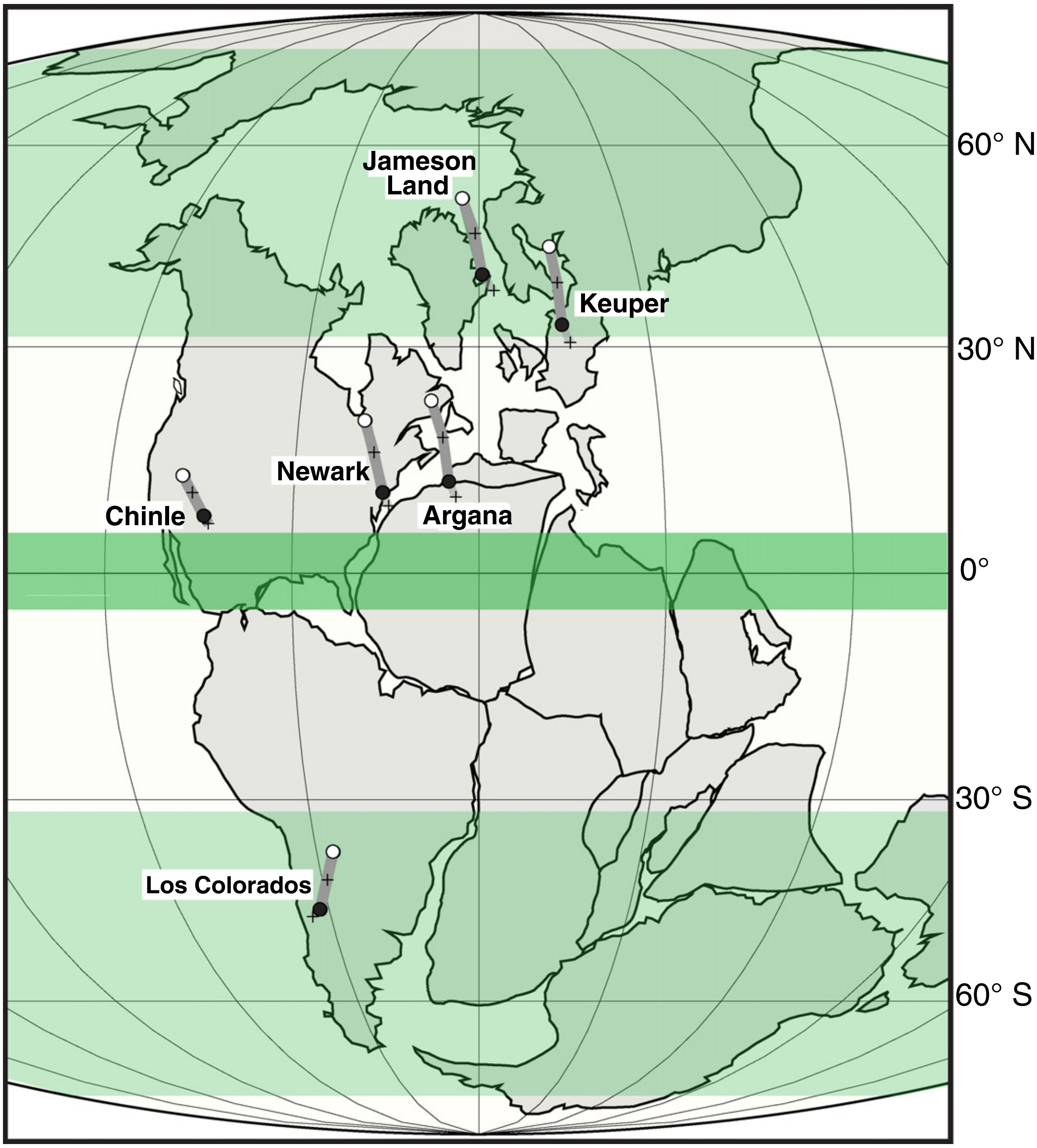


Fig. 18. Paleogeography of the Late Triassic world. Reconstruction of Pangea is based on continental rotation parameters of Lottes and Rowley (1990) and positioned in latitude using a 220 Ma mean global pole from Kent and Irving (2010). Some key continental localities are indicated by filled circles (for positions at 220 Ma) connected to their relative positions at 200 Ma by open circles, with relative positions of localities at 230 Ma and 210 Ma indicated by crosses. Modeled zonal belts (Manabe and Bryan, 1985) of precipitation (P) relative to evaporation (E) are indicated by darker and medium green shading for P > E (more humid) and lighter shading for P < E (more arid). Figure adapted from Kent et al. (2014).

Washington Valley Member and k = 498.195 (interpolated) for the base of Chron E23r.

- The portion of the Newark-Hartford APTS that has direct astrochronologic control (Chrons E8n to H27n, k = 554.882 to 492.186, age range 224.54 to 199.15 Ma) is over 25 Myr long, making it one of the longest continuous astrochronostratigraphic polarity time-scales. For example, the Oligocene portion of the Cenozoic APTS is 13 Myr-long (Palike et al., 2006) whereas the rest of the Cenozoic APTS is assembled from shorter segments. However, the astrochronological element of the Newark-Hartford APTS is exceeded in continuous length of record by the 70 Myr-long Triassic-Jurassic Inuyama deep-sea bedded chert sequence of Japan (Ikeda and Tada, 2014) whereas the polarity sequence of the Newark-Hartford APTS is far exceeded in length of record by the 100 Myr (Barremian to Oligocene) geomagnetic polarity history recorded from magnetostratigraphy of the Umbrian Apennines (Lowrie and Alvarez, 1981) and the 180 Myr-long marine magnetic anomaly record of geomagnetic polarity reversals (reviewed by Gee and Kent (2007).
- Triassic stage and substage divisions have classically been erected using ammonoid and/or conodont biostratigraphy from the Tethyan marine realm, often from type areas in the Alps with floating and vague chronologies. Magnetostratigraphic correlation of key Tethyan sections with diagnostic biostratigraphic elements to the Newark-Hartford APTS allows determination of numerical ages of standard marine stages, as follows: 227 Ma for the Carnian/Norian boundary, 205.5 Ma for the Norian/Rhaetian boundary (using a chemostratigraphic criterion, or about 4 Myr older for alternative criteria), and 199.5 Ma for the Hettangian/Sinemurian boundary. The Triassic/Jurassic boundary projects into the Newark-Hartford APTS at 201.4 Ma from the St. Audrie's Bay section. The Triassic/Jurassic boundary, as presently defined, thus post-dates the end-Triassic extinction event by about 200 kyr.
- The underlying basis of the Newark-Hartford astrochronology was strongly corroborated by high-precision U-Pb dating of CAMP lavas that bracket cyclic sedimentary units (Blackburn et al., 2013). There is close agreement between relative time from astrochronology and from the high-precision U-Pb dates over the 600 kyr CAMP interval, indicating that orbital signals are faithfully recorded in these lacustrine strata and that solar system dynamics along with the duration of the

long eccentricity cycles can be reliably modeled that far back. The good agreement between the 230.91 ± 0.33 Ma U-Pb zircon age from an ash layer in late Carnian marine sediments (Furin et al., 2006) with correlatives linked to the Newark-Hartford APTS further validates its ability to apportion time using orbital climate cyclicity over a 30 Myr scale.

- CAMP activity was recognized as representing episodic volcanism over several million years at around 200 Ma based on compilations of $^{40}Ar/^{39}Ar$ dates (e.g., Fig. 6 in Knight et al. (2004)). However, high-precision U-Pb zircon dating now strongly places CAMP volcanism starting closer to 201.6 Ma and, importantly for its potential environmental impact(s), only about 600 kyr in duration (Blackburn et al., 2013). The $^{40}Ar/^{39}Ar$ technique is gradually being normalized with respect to astronomical cycles (Kuiper et al., 2008) and intercalibrated with U-Pb ages (Renne et al., 2010), which should reduce systematic bias arising from comparing ages from different isotopic systems. As one reviewer suggested, some caution should nevertheless be exercised with respect to the high-precision U-Pb ages that are used to anchor the stratigraphic column. The dispersions of high-precision U-Pb dates in some of the cited studies (e.g., Blackburn et al., 2013) are often greater than predicted from analytical uncertainty and may be real, reflecting time-scales of zircon crystallization on the 10–100 kyr level (Caricchi et al., 2014; Ickert et al., 2015). Unresolved open system behavior can also bias ages at the sub-permil level even with the use of chemical abrasion (Mattinson, 2005).
- Suggestions of a cryptic unconformity with a 3 Myr or longer missing interval of Rhaetian time in the Newark basin are incompatible with the recorded presence of Chron E23r immediately below the oldest CAMP basalts in three different basins, which would be totally unexpected from uplift and erosion or nondeposition on the required regional scale. Moreover, magnetostratigraphic correlation of strata with demonstrable Rhaetian affinities at St. Audrie's Bay (Hounslow et al., 2004) to the Newark-Hartford APTS shows there is no evidence for a missing Rhaetian in the Newark Supergroup.
- Paleosol carbonates in the Newark and Hartford basins are an excellent proxy for generating a record of atmospheric pCO_2 concentrations that can be registered directly to the Newark-Hartford APTS (Schaller et al., 2015). Tethyan marine carbonate sections with magnetobiostratigraphic constraints for correlation to the Newark-Hartford APTS provide a complementary record of carbon cycling from measurements of bulk carbonate $\delta^{13}C$ (Muttoni et al., 2014). The last few million years of Late Triassic time showed a steady decrease in pCO_2 levels that may be due to increasing chemical weathering across the equatorial belt with northward drift of the Pangea landmass. This same interval was also characterized by highly fluctuating $\delta^{13}C$ values whose origin is unclear.
- The latitudinal gradient in climate at any given time – from polar night, snow fields and even ice sheets to temperate climates, tropical deserts and the equatorial humid belts – probably exceeds climate change at any given place over geological spans of time. This emphasizes the need for differentiating global climate change from changes or differences in latitudinal setting where comparisons and observations are made, especially of terrestrial fauna and flora. The missing Rhaetian assertion, based on apparent temporal delays in appearances and disappearances of palynoflora, conchostracans, and other endemic taxa, is an example in our view of the miscorrelation of ecostratigraphy as chronostratigraphy.

Acknowledgements

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