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The Late Triassic timescale: Age and correlation of the Carnian–Norian boundary

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Abstract

The Late Triassic timescale is poorly constrained due largely to the dearth of reliable radio-isotopic ages that can be related precisely to biostratigraphy combined with evident contradictions between bio-stratigraphic and magnetostratigraphic correlations. These problems are most apparent with regard to the age and correlation of the Carnian–Norian boundary (base of the Norian Stage). We review the available age data pertaining to the Carnian–Norian boundary and conclude that the "long Norian" in current use by many workers, which places the Carnian–Norian boundary at ~228 Ma, is incorrect. The evidence supports a Norian stage that is much shorter than proposed by these workers, so the Carnian–Norian boundary is considerably younger than this, close to 220 Ma in age. Critical to this conclusion is the correlation of the Carnian–Norian boundary in nonmarine strata of Europe and North America, and its integration with existing radioisotopic ages and magnet-ostratigraphy. Three bio-stratigraphic datasets (palynomorphs, conchostracans and tetra-pods) reliably identify the same position for the Carnian–Norian boundary (within normal limits of bio-stratigraphic resolution) in nonmarine strata of the Chinle Group (American Southwest), Newark Supergroup (eastern USA–Canada) and the German Keuper. These biostratigraphic datasets place the Carnian–Norian boundary at the base of the Warford Member of the lower Passaic Formation in the Newark Basin, and, as was widely accepted prior to 2002, this correlates the base of the Norian to a horizon within Newark magnet-ozone E13n. In recent years a correlation based solely on magnetostratigraphy has been proposed between the Pizzo Mondello section in Sicily and the Newark section. This correlation, which ignores robust biostrati-graphic data, places the Norian base much too low in the Newark Basin section (~at the base of the Lockatong Formation), correlative to a horizon near the base of Newark magnet-ozone E8. Despite the fact that this correlation is falsifiable on the basis of the bio-stratigraphic data, it still became the primary justification for placing the Carnian–Norian boundary at ~228 Ma (based on Newark cyclo-stratigraphy). The "long Norian" created thereby is unsupported by either bio-stratigraphic or reliable radioisotopic data and therefore must be abandoned. While few data can be presented to support a Carnian–Norian boundary as old as 228 Ma, existing radio-isotopic age data are consistent with a Norian base at ~220 Ma. Although this date is approximately correct, more reliable and precise radio-isotopic ages still are needed to firmly assign a precise age to the Carnian–Norian boundary.

1. Introduction

During the last 30 years, remarkable progress has been made in defining and refining a global timescale for the Triassic Period, particularly due to the efforts of the Subcommission on Triassic Stratigraphy (STS) of the IUGS International Commission on Stratigraphy (Lucas, 2010a, 2010b). Traditionally, the biostratigraphy of conodonts and ammonoids has provided the foundation of the Triassic timescale (e.g., Mojsisovics, 1882a, 1882b; Mojsisovics et al., 1895; Mojsisovics, 1902; Tozer, 1967, 1971, 1974, 1984; Kozur, 1980; Krystyn, 1991; Kozur, 2003; Krystyn, 2008; Balini et al., 2010; Orchard, 2010), supplemented by other marine fossils. Most notable of these are radiolarians and bivalves (e.g., Dumitrică, 1978a, 1978b; Kozur and Mostler, 1979; Nakaseko and Nishimura, 1979; Pessagno et al., 1979; Dumitrică et al., 1980; Kozur and Mostler, 1981; Dumitrică, 1982a, 1982b, 1982c; Yao, 1982; Yao et al., 1982; Yao, 1990; Yeh, 1990; Carter, 1993; Kozur and Mostler, 1994, 1996; Tekin, 1999; DeWever et al., 2001; McRoberts, 2010; O'Dogherty et al., 2010), but there is a growing contribution from some nonmarine fossil groups, such as palynomorphs, conchostracans and tetrapod vertebrates (e.g., Schulz, 1962; Mädler, 1964; Schulz, 1965; Scheuring, 1970; Chang et al., 1976; Visscher and Brugman, 1981; Kozur and Seidel, 1983a, 1983b; Orłowska-Zwolińska, 1985; Lucas, 1998, 1999; Kozur and Weems, 2005; Schulz and Heunisch, 2005; Kozur and Weems, 2007; Lucas, 2010c; Kozur and Weems, 2010a; Cirilli, 2010; Kuerschner and Herngreen, 2010). The integration of Triassic biostratigraphy with radioisotopic ages and magnetostratigraphy also has advanced significantly during the past three decades (e.g., Szurlies, 2001; Bachmann and Kozur, 2004; Hounslow et al., 2004; Kozur and Bachmann, 2008; Ogg et al., 2008; Kozur and Bachmann, 2010b; Hounslow and Muttoni, 2010; Mundil et al., 2010).

Even so, serious problems remain with regard to the Late Triassic timescale due to a dearth of reliable radioisotopic ages that can be related precisely to biostratigraphy and also to some evident contradictions between biostratigraphic and magnetostratigraphic correlations. These problems are most apparent with regard to the age and correlation of the Carnian–Norian boundary (base of the Norian Stage). Here, we review the age data that pertain to the Carnian–Norian boundary and conclude that the "long Norian" in current use by many workers, which places the Carnian–Norian at ~228 or 229 Ma (e.g. Krystyn et al., 2002; Gallet et al., 2003; Muttoni et al., 2004) lacks robust support and that the Carnian–Norian boundary is much younger, close to 220 Ma. Critical to this conclusion is the correlation of the Carnian–Norian boundary in nonmarine strata of Europe and North America (Fig. 1), and its integration with existing radioisotopic ages and magnetostratigraphy.

2. Carnian–Norian boundary in the marine section

At present, no Global Stratotype Section and Point (GSSP) has been established to define the base of the Norian Stage. The working definition for the Norian base has long been the base of the *Stikinoceras kerri* ammonoid zone in the North American succession (e.g., Silberling and Tozer, 1968; Tozer, 1994; Krystyn et al., 2002; Kozur, 2003; Ogg, 2004; Balini et al., 2010) (Fig. 2). In the Tethyan realm, this level is approximately equivalent to the boundary between the *Anatropites spinosus* and *Guembelites jandianus* ammonoid zones (Krystyn, 1980; Krystyn et al., 2002; Ogg, 2004; Balini et al., 2010). The STS appears likely to propose to define the Carnian–Norian boundary formally by a conodont datum at one of two proposed GSSP locations: Black Bear Ridge on Williston Lake in British Columbia, western Canada, or Pizzo Mondello in Sicily, Italy (e. g., Muttoni et al., 2004; Orchard, 2007; Nicora et al., 2007).

Orchard (2010) has provided the most recent summary of the current status of a conodont-defined Norian base, noting that at a combination of different sections in western Canada there is a good correspondence between ammonoid, bivalve, and conodont faunal changes at, or close to, the traditional boundary level (Orchard et al., 2001; McRoberts, 2007, 2010). However, the North American Norian conodont succession contains several endemic forms that are not present in the Eurasian Tethys. Moreover, although Black Bear Ridge has a good conodont and halobiid bivalve record across the Carnian–Norian boundary, it has no radiolarians, relatively few
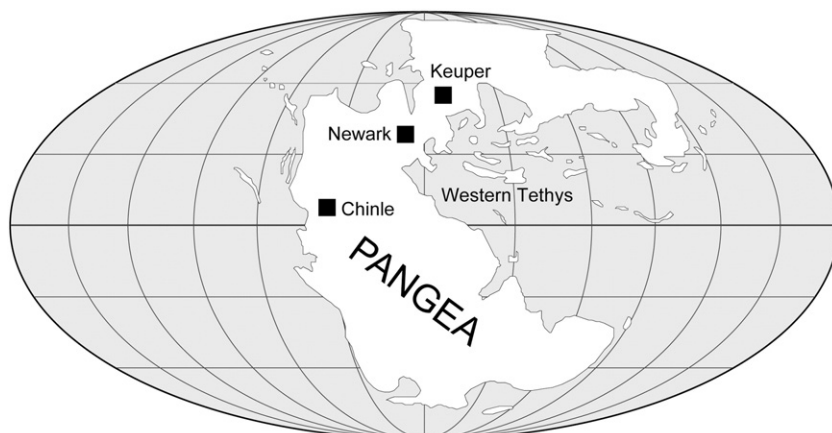

Fig. 1. Map of Late Triassic Pangaea (from McRoberts, 2010) showing locations of Chinle Group, Newark Supergroup and Keuper sections discussed in text. The marine Carnian–Norian boundary sections (including Pizzo Mondello) discussed in the text were located in western Tethys.

ammonoids, and the CAI (conodont alteration index) is 5, which precludes magnetostratigraphy and reliable palynological investigations. Because of these limitations, some view Pizzo Mondello (western Sicily, Italy) as much more suitable as a GSSP for the base of the Norian. Ammonoids are relatively uncommon in this section as well, as they are in most uncondensed Norian sections in the world, but halobiids and radiolarians are common. The rocks are unaltered and have allowed the establishment of a detailed magnetostratigraphy (Muttoni et al., 2004, 2010; Balini et al., 2010). Detailed palynological studies have not yet been undertaken. Because of the detailed magnetostratigraphy, the Pizzo Mondello section has a high potential for correlation with continental successions. Also, the correlation potential for the marine facies is better than for the Black Bear Ridge section, because Pizzo Mondello has a typical Tethyan conodont fauna, the species of which are much more widely distributed than the partly endemic species from the western margin of the North American continent (e.g., Orchard, 2010).

In regard to the question of the long versus short Norian, the few uncertainties remaining as to the final GSSP and the final position of the Norian base are only of minor concern. The different proposals for the base of the Norian, e.g., base of the *G. jandianus* Zone in the Tethys, base of the *S. kerri* Zone in western North America, FAD of *Halobia austriaca* and *H. beyrichi* in the Tethys and in western North America, or the variously proposed conodont datums, such as the FAD of *Epigondolella*/*Carnepigondolella orchardi* Kozur, FAD of *Carnepigondolella pseudoechinata* (Kozur), FAD of *Norigondolella navicula* (Huckriede), or the FAD of *Epigondolella quadrata* Orchard, do not differ by a sufficiently substantial time interval (probably less than one million years) to change the outcome of this problem (Fig. 2). So, for the purposes of our discussion, we will regard the biostratigraphic base of the Norian as the base of the *S. kerri* Zone in western North America, which is its traditional position.

3. Long Norian vs. short Norian

Prior to the mid-1990s, estimates of the age of the base of the Norian Stage hovered around 220 Ma (Fig. 3). In the 1980s, these estimates were based loosely on K–Ar ages of intrusives, such as the Indonesian tin granites (average age 219 ± 4 Ma), which cut rocks as young as Norian, or on the K–Ar ages of basalt flows, such as the Sugars Basalt (average age 230 ± 7 Ma), which underlies palynologically-dated Carnian strata (see review by Forster and Warrington, 1985). Unfortunately, these radioisotopic ages not only were based on the relatively imprecise K–Ar method, but they also lacked any precise relationship to robust biostratigraphy.

In the 1990s, Olsen and Kent (1996) and Kent and Olsen (1999) advocated counting Milankovitch-forced cycles of deposition in the Newark Supergroup section of New Jersey (USA) to estimate the age of the Carnian–Norian boundary (Fig. 4). They reasoned that identification and counting of the number of Milankovitch-frequency sedimentary cycles between the oldest radioisotopically-dated basalt and the palynologically-defined Carnian–Norian boundary in the Newark Basin should yield a reasonable estimate for the age of the base of the Norian. Olsen and Kent (1996) inferred that the primary cyclostratigraphic unit in the Newark Basin (called a McLaughlin cycle) corresponds to the long eccentricity cycle, with a duration of 400 ka. Using a date of ca. 202 Ma for the age of the Orange Mountain Basalt, the 36 to 38 cycles below the basalt (to the palynological boundary) yields an age of ca. 217–218 Ma for the Norian base (cf. Olsen et al., 2002). This estimate was widely used, notably appearing in the compilation of Ogg (2004), which represented the then-agreed-upon Triassic timescale of the International Commission on Stratigraphy (Fig. 3).

However, Krystyn et al. (2002), Gallet et al. (2003) and Muttoni et al. (2004) advocated a much older Norian base based on their proposed magnetostratigraphic correlation of marine Carnian–Norian boundary sections to the Newark Supergroup section. The magnetostratigraphic correlation that they created placed the Carnian–Norian boundary much lower in the Newark section than earlier correlations and well below the long-accepted palynostratigraphic Carnian–Norian boundary. Using cycle counting, the age of this lower position of the Norian base was estimated to be ca. 228 Ma (Fig. 4). Likewise, based on a correlation from marine sections to the Newark Supergroup by magnetostratigraphy, Channell et al. (2003) recognized a slightly shorter Norian with an assumed base at 226 Ma.

Thus arose what Muttoni et al. (2004) referred to as the "long Norian." With a base calibrated at 228 Ma, and a Rhaetian base estimated at about 204–205 Ma (Ogg et al., 2008), the duration of the "long Norian" dwarfs that of any other Triassic stage and represents almost half the duration of the entire Triassic Period (Ogg et al., 2008, p. 102). Significantly, the concept of the "long Norian" has been widely and uncritically accepted and incorporated into timescales used by geological societies (Walker and Geissman, 2009) and used to estimate the ages of everything from Late Triassic shark egg capsules (Fischer et al., 2010) to the tempo and mode of early dinosaur evolution (Langer et al., 2010).

Unfortunately, the rapid and widespread acceptance of a "long Norian" ignored the fact that it was based on a very tentative magnetostratigraphic correlation that repudiated well-founded and long accepted biostratigraphic placement of the Carnian–Norian boundary in
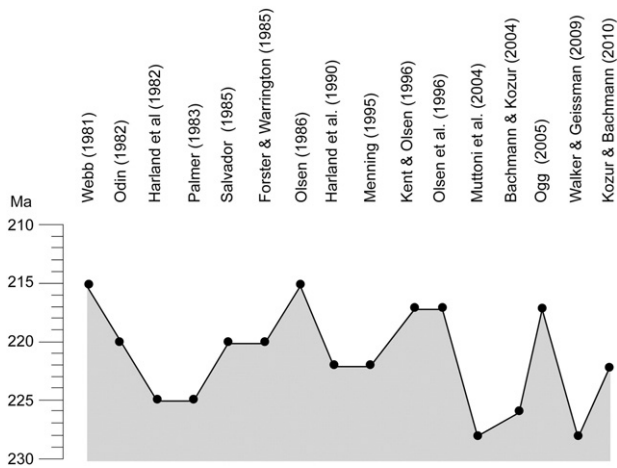
Stage	Substage	Ammonoid zones — American Cordillera	Ammonoid zones — Tethyan	Conodonts (North America)
Norian	Sevatian	Gnomohalorites cordilleranus	Sagenites quinque-punctatus	Cypridodella bidentata
Norian	Alaunian	Mesohimavatites columbianus	Halorites macer	Cypridodella serrulata
Norian	Alaunian	Mesohimavatites columbianus	Halorites macer	Cypridodella postera
Norian	Alaunian	Mesohimavatites columbianus	Himavatites hogarti	Cypridodella spiculata
Norian	Alaunian	Drepanites rutherfordi	Cyrtopleurites bicrenatus	Cypridodella multidentata
Norian	Lacian	Juvavites magnus	Juvavites magnus	Epigondolella triangularus
Norian	Lacian	Malayites dawsoni	Malayites paulckei	Epigondolella quadrata
Norian	Lacian	Stikinoceras kerri	Guembelites jandianus	
Carnian	Tuvalian	Klamathites macrolobatus	Anatropites spinosus	Metapolygnathus primitius
Carnian	Tuvalian	Tropites welleri	Tropites subbullatus	Metapolygnathus nodosus / Metapolygnathus carpathicus
Carnian	Tuvalian	Tropites dilleri	Tropites dilleri	Metapolygnathus polygnathiformis
Carnian	Julian	Sirenites nanseni Austrotrachyceras obesum	Austrotrachyceras austriacum	Metapolygnathus tadpole
Carnian	Julian	Trachyceras desatoyense	Trachyceras aonoides / Trachyceras aon	Metapolygnathus tadpole
Carnian	Julian	Daxatina canadensis	Daxatina canadensis	Metapolygnathus intermedius

Fig. 2. Marine biostratigraphy of the Carnian–Norian boundary (after Lucas, 2010a).

Fig. 3. Comparison of numerical age estimates of the Carnian–Norian boundary during the last 30 years.

the Newark Supergroup at or near the base of the Passaic Formation in the New Jersey section (Fig. 4). In other words, the Carnian–Norian boundary in the Newark Supergroup based on palynostratigraphy and vertebrate biostratigraphy was rejected by Krystyn et al. (2002), Gallet et al. (2003) and Muttoni et al. (2004) in favor of their correlation of the magnetostratigraphy of Turkish sections and the Pizzo Mondello section to the Newark magnetostratigraphy. Here, we argue that this rejection was unwarranted because three quite different biostratigraphic datasets (palynomorphs, vertebrates and conchostracans) place the Norian base near the base of the Passaic Formation in the Newark section. This biostratigraphic datum thus requires a different magnetostratigraphic correlation than Krystyn et al. (2002), Gallet et al. (2003) and Muttoni et al. (2004) advocated, and therefore eliminates the basis for the "long Norian" and also for the "moderately long Norian" of Channell et al. (2003). Even so, problems still remain with precisely estimating the numerical age of the Norian base largely because of unresolved contradictions among published radioisotopic ages.

4. Carnian–Norian boundary in the nonmarine section

4.1. Introduction

Placement of the Carnian–Norian boundary in nonmarine strata can be achieved using three biostratigraphic datasets: palynomorphs, conchostracans and tetrapod vertebrates. These datasets agree well in their placement of the boundary in three critical sections: the German Keuper (Fig. 5), Newark Supergroup of the eastern USA (Fig. 6) and the Chinle Group of the American Southwest (Fig. 7). Here, we review these biostratigraphic datasets and explain their bearing on the placement of the Carnian–Norian boundary in Upper Triassic nonmarine strata.

4.2. Palynomorphs

Biostratigraphic placement of the Carnian–Norian boundary by palynomorphs in nonmarine strata of Europe and North America began in the 1970s with recognition that marine strata of the Alpine Triassic yield distinctive palynomorph assemblages that can also be recognized in nonmarine strata (e.g., Dunay and Fisher, 1974; Visscher and Brugman, 1981). The taxon key to these correlations is *Camerosporites secatus*, which has a well-established Ladinian–Carnian stratigraphic range in the marine Alpine section (Dunay and Fisher, 1978; Visscher and Brugman, 1981; Van der Eem, 1983; Blendinger, 1988). Indeed, Visscher and Krystyn (1978) introduced the concept of a Carnian "*C. secatus* phase," characterized by a rapid diversification of circumpolloid taxa such as *C. secatus*, *Duplicisporites granulatus* and

Praecirculina granifer, associated with a group of monosaccate spores (*Enzonalasporites vigens*, *Patinasporites densus*, *Pseudoenzonalasporites summus* and *Vallasporites ignacii*) and the bisaccate *Samaropollenites speciosus*. This phase was intended to provide a global palynostratigraphic correlation because its characteristic assemblages are widely distributed across Late Triassic Pangaea, being known from localities in Europe, North Africa, Israel, Timor, Australia, the U.S.A. and Arctic Canada (see review by Cirilli, 2010). Significantly, several palynomorph records of the *Camerosporites secatus* phase come from strata correlated by ammonoids and/or conodonts to the Late Triassic marine timescale. A pivotal point is that *C. secatus* is known only from Ladinian and Carnian strata, whereas its inferred descendant, *C. verrucosus*, is known only from Norian strata (Litwin and Skogg, 1991).

An important example of the stratigraphic range of *C. secatus* is in the Germanic Basin, where Herngreen (2005) and Kuerschner and Herngreen (2010) recently defined a *C. secatus* Zone (Fig. 5). They defined the base of this zone by the FO (first occurrence) of *C. secatus*, which coincides with the FOs of *E. vigens*, *Triadispora verrucata* and *V. ignacii*, and the first common occurrence of *Ovalipollis pseudoalatus*. The zone is found in the upper Grabfeld Formation through the Mainhardt Formation (Lower Gipskeuper to lower Steinmergelkeuper),
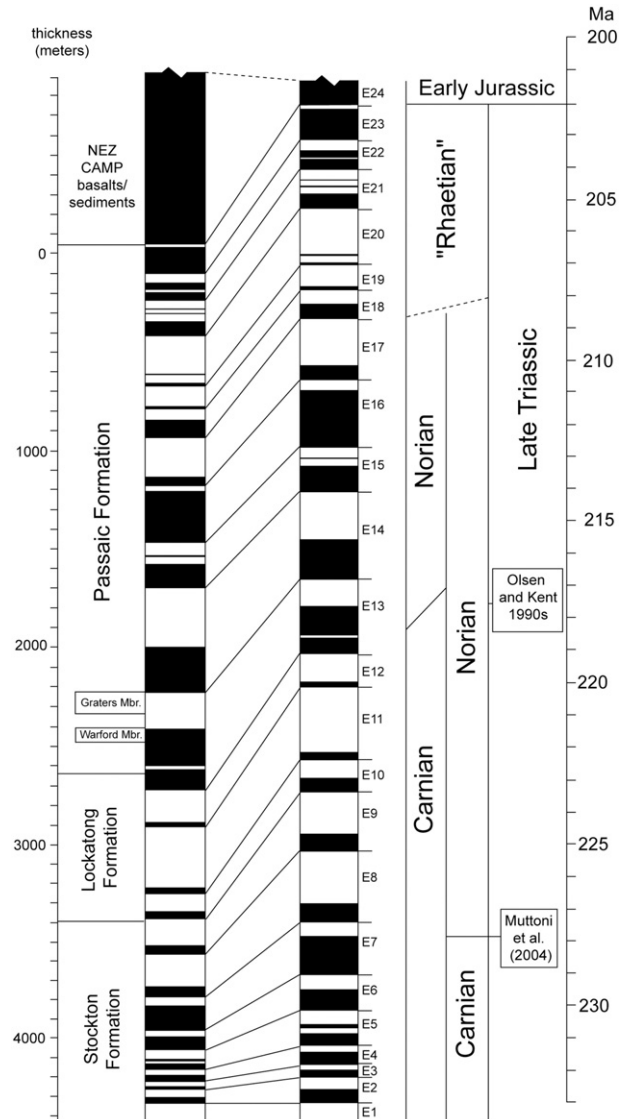

Fig. 4. Magnetostratigraphic zonation of the Newark Basin section (after Kent and Olsen, 1999) showing the two inferred positions of the Carnian–Norian boundary.

LITHOSTRATIGRAPHY			Palynomorphs	Conchostracans	Tetrapods	Age	
Triletes Beds		Exter Fm.	Rhaetipollis germanicus	Euestheria brodeiana	Apachean	Rhaetian	Norian
Contorta Beds				Gregoriusella polonica			
Trossingen Fm. (Knollenmergel)	Postera Sandstone						
Lowenstein Fm. (Stubensandstein 4)	Arnstadt Fm. — Postera Beds (lower/middle)					Sevatian	
Lowenstein Formation (part) Stubensandstein 1-3, Hangendletten 1-3, Obere Bunte Mergel (part)	Arnstadt Formation		Granuloperculatipollis rudis	Shipingia olseni	Revueltian	Alaunian	
				Shipingia hebaozhaiensis			
				N. barnaschi- S. mcdon.			
				small Shipingia and large Euestheria		Lacian	
				Euestheria buravasi- Euestheria n. sp.			
				Palaeolimnadia schwanbergensis zone			
Mainhardt Fm. (Heldburggips./Obere Bunte Megel[part])	Weser Formation (Oberer Gipskeuper)		Camerosporites secatus	L. freybergeri-P. schwanbergensis	Adamanian	Tuvalian	Carnian
Hassberge Fm. (Coburg Ss/Blasensandstein)				Laxitextella freybergeri			
Steigerwald Formation — Lehrberg Beds				Laxitextella seegisi			
Rote Wand				Eosolimnadiopsis gallegoi			
Stuttgart Formation — Dunkle Mergel — Schilfsandstein				Schilfsandstein faunas	Otischalkian	Julian	
upper Grabfeld Formation ("Estherienschichten")				Laxitexella laxitexta			
				Laxitexella mutlireticulata			

Fig. 5. Biostratigraphy of the Carnian–Norian boundary interval in the Keuper section of the Germanic Basin. Based primarily on Kozur and Bachmann (2008, 2010b), Kozur and Weems (2010a), Kuerschner and Herngreen (2010) and Lucas (2010c).

strata assigned a Carnian age based on diverse criteria (Bachmann and Kozur, 2004; Kozur and Bachmann, 2005, 2008) (Fig. 5). The Carnian can be further divided palynostratigraphically, with an assemblage containing *Paracirculina quadruplicis* in addition to the common Carnian forms *C. secatus*, *V. ignacii* and *P. densus* characterizing the transition from the Julian to the Tuvalian (Cirilli, 2010).

LITHOSTRATIGRAPHY	Palynomorphs	Conchostracans	Tetrapods	Age	
Newark extrusive zone (Minden Group)	Classopollis meyeriana zone	Bulbilimnadia froelichi zone	Wassonian LVF	Hettangian	JURASSIC
		Bulbilimnadia sheni zone			
		Bulbilimnadia killianorum zone			
		uppermost Euestheria brodeiana zone		Rhaetian	LATE TRIASSIC
Passaic Formation	Passaic palynological event — upper Balls Bluff/upper Passaic		Cliftonian LVF		
		Shipingia olseni zone		Norian	
		Redondaestheria grovetonensis			
	transition zone	Shipingia hebaozhaiensis			
	lower Passaic/ Heidelsburg	Shipingia mcdonaldi	Neshanician LVF		
		small Shipingia and large Euestheria			
		Euestheria buravasi- Euestheria n. sp.			
	transition zone	Palaeolimnadia schwanbergensis	Revueltian LVF / Adamanian LVF		
Lockatong Formation		Wanneresheria pennsylvanica zone	Conewagian LVF	Carnian	
	New Oxford/ Lockatong	Howellisaura ?ovata			
		Howellisaura princetonensis zone			
Stockton Formation		undefined conchostracan zone	Sanfordian LVF		
		Laxitextella multireticulatus zone?			

Fig. 6. Biostratigraphy of the Carnian–Norian boundary interval in the Newark Supergroup section. Based primarily on Litwin et al. (1991), Huber et al. (1993), Lucas and Tanner (2007) and Kozur and Weems (2010a).

Fig. 7. Biostratigraphy of the Carnian–Norian boundary interval in the Chinle Group section. Based primarily on Litwin et al. (1991), Lucas (1993, 2010c) and Kozur and Weems (2010a).

In the palynostratigraphy of Kuerschner and Herngreen (2010), the succeeding zone is the *Granuloperculatipollis rudis* Zone (Fig. 5), which lacks *C. secatus* and has its base at the FO of common *G. rudis*. In this zone, the circumpolles *Classopollis meyeriana*, *C. zwolinskae* and *G. rudis* become common. In the Germanic Basin, this zone begins the Heldburg Gypsum Member and correlatives (Orłowska-Zwolińska, 1983; Heunisch, 2005; Schulz and Heunisch, 2005) of late Tuvalian age (Bachmann and Kozur, 2004) and is present in the late Tuvalian part of the marine Opponitz Formation of the Alps (Roghi et al., 2010). The zone is widely known from the Löwenstein Formation and correlatives (Arnstadt Formation, Steinmergelkeuper: Schulz and Heunisch, 2005), which are strata assigned a Norian age based on diverse criteria (Bachmann and Kozur, 2004; Kozur and Bachmann, 2005, 2008).

In North America, Upper Triassic palynomorph assemblages are known from the Chinle Group strata of the American Southwest (e.g., Dunay and Fisher, 1979; Fisher and Dunay, 1984; Litwin et al., 1991; Cornet, 1993) and from the Newark Supergroup strata of the eastern USA and Canada (e. g., Dunay and Fisher, 1974; Cornet and Traverse, 1975; Cornet, 1977; Cornet and Olsen, 1985; Manspeizer and Cousminer, 1988; Robbins et al., 1988; Litwin and Ash, 1993). Litwin et al. (1991) provided the most recent analysis of the Chinle Group palynoflora, organizing it into three zones (Fig. 7). Zone II includes the most diverse and abundant palynological assemblages in the Chinle Group, and is characterized by the FOs of *C. rudis*, *E. vigens*, *Heliosaccus dimorphus*, *Ovalipollis ovalis*, *P. summus* and other taxa (e.g. *Alisporites* spp., *Cycadopites stonei*, *Guthoerlisporites cancellosus*), and the HOs (highest occurrences) of *Brodispora striata*, *C. secatus*, *Equisetosporites chinleanus* and *Lagenella martini*. This assemblage closely resembles upper Carnian assemblages from Europe, so Litwin et al. (1991) assigned it a Tuvalian age (Fig. 7). Indeed, the HOs of *C. secatus*, *Triletes klausii* and *B. striata* confirm that the zone II assemblages are no younger than late Carnian. Founded on the FOs of *Camerosporites verrucosus* and *Kyrtomisporis* spp., zone III was assigned to an early Norian age because of the absence of significant Carnian taxa (e.g. *B. striata* and *C. secatus*) and the presence of *C. verrucosus* and *P. summus*. Based on this palynostratigraphic zonation, the Norian base in the Chinle Group is close to the base of the Sonsela Member of the Petrified Forest Formation and its correlatives (Fig. 7).

The palynomorph record across the Carnian–Norian boundary in the Newark Supergroup has been documented primarily by Cornet (1977), Cornet and Olsen (1985), Ediger (1986), Traverse (1987), Robbins and Weems (1988) and Fowell (1994). Key Newark palynomorph assemblages come from the Fundy Basin in Nova Scotia, the Newark Basin in New Jersey, the Gettysburg Basin in Pennsylvania, and the Dan River-Danville, Deep River and Richmond Basins in North Carolina and Virginia (Litwin et al., 1991). Here, four successive palynological zones were established by Cornet (1977) and Cornet and Olsen (1985), and are now referred to by a terminology slightly modified by Litwin et al. (1991) (Fig. 6). Within this zonation, the HO of *C. secatus* is in the New Oxford–Lockatong palynological zone, so the Carnian–Norian boundary falls between the New Oxford–Lockatong and the Lower Passaic–Heidlersburg palynofloral zones (Fig. 6). Furthermore, the FO of the Norian indicator *C. veruccosus* lies in the Graters Member of the Passaic Formation (Cornet, 1977; Litwin and Skogg, 1991). In the Newark Basin, palynofloras of the New Oxford–Lockatong palynofloral zone are found in the lower part of the Passaic Formation, whereas the stratigraphically lowest palynofloras of the Lower Passaic–Heidlersburg palynofloral zone are from the lower to middle part of the Passaic Formation (Figs. 4 and 6). Palynostratigraphy thus constrains the position of the Carnian–Norian boundary to the lower part of the Passaic Formation (e.g., Kent and Olsen, 1999, Fig. 4) (Figs. 4 and 6).

Correlation of the Chinle palynomorph record to the Newark Supergroup palynomorphs and placement of the Carnian–Norian boundary in the Chinle Group and the Newark Supergroup based on palynomorphs is therefore relatively straightforward and well accepted (Dunay and Fisher, 1974; Litwin et al., 1991; Cornet, 1993; Fowell, 1994) (Fig. 8). It is also reinforced by megafossil plant correlations between the Chinle and Newark (Ash, 1980, 1987). Thus, the bulk of the lower Chinle palaeofloras are in the *Dinophyton* floral zone of Ash (1980), which corresponds to the Chinle zone II palynoflora (Fig. 7) and is found in the Newark Supergroup in the middle part of the New Oxford Formation (Gettysburg Basin), uppermost Stockton Formation (Newark Basin), Cumnock Formation (Deep River Basin) and Cow Branch Formation (Dan River Basin) (e.g., Ash, 1980; Cornet and Olsen, 1985; Axsmith and Kroehler, 1988; Lucas and Huber, 1993, 2003). Therefore, the megaflora equates most of the lower Chinle Group (strata below the HO of *C. secatus*) to these Newark Supergroup formations, all of which are older than the Passaic Formation in the Newark Basin, reinforcing the palynostratigraphic correlation (Fig. 8).

4.3. Conchostracans

Conchostracans are bivalved crustaceans that live in freshwater lakes and ponds. Their minute, drought-resistant eggs can be dispersed by the wind, and this guaranteed a broad geographic range to some conchostracan taxa across much of Triassic Pangaea. Triassic conchostracan biostratigraphy has been developed by several workers, most recently by Kozur and Weems (2005, 2007, 2010a) in Europe and North America.

In the Germanic Basin, the late Tuvalian has a rich conchostracan fauna (Fig. 5), which is, however, restricted to some horizons of the basin margins, as the central basinal hypersaline deposits do not contain conchostracans. Most conchostracans occur in strata deposited during the time of marine ingressions, when they lived in areas where rivers transported fresh water into the basin. Conchostracans, triopsids and other fossils can be found, sometimes in an excellent state of preservation in mudstone and siltstone intercalations in basin marginal sandstone bodies that represent abandoned channels, deltaic plains or small ponds on flood plains close to the margin of the basin.

In the Germanic Basin, a marine ingression can be observed at the base of the predominantly hypersaline Weser Formation and its correlatives (Fig. 5). The Dolomie de Beaumont with marine bivalves was deposited during this ingression in the southwestern Germanic

selected index taxa	Age		Chinle Group	Newark Supergroup	Keuper
● FO Palaeolimnadia; ● FO Camerosporites verrucosus; ● FO Aetosaurus	Rhaetian	Apachean	Rock Point Formation	Orange Mtn. Basalt	Exter Formation
	Norian	Revueltian	Owl Rock Formation; Painted Desert Member; Sonsela Member (Petrified Forest Formation)	Passaic Formation	Arnstadt Formation
● HO Laxitextella; ● HO Camerosporites secatus	Carnian	Adamanian	Blue Mesa Member; Bluewater Creek Formation (Petrified Forest Formation)	Lockatong Formation	Mainhardt Fm.; Hassberge Fm.; Steigerwald Fm.
		Otischalkian	Shinarump Formation	Stockton Formation	Stuttgart Formation

Fig. 8. Nonmarine correlation of the Carnian–Norian boundary based on Figs. 5–7.

Basin (eastern France and southwestern southern Germany). These marine bivalves, e.g., *Costatoria vestita* (von Alberti), indicate an early Tuvalian age. The Dolomie de Beaumont changes across most of southern Germany and the Thuringian Basin into an estuarine area (Kozur and Bachmann, 2010a) that in southern Germany and eastern France contains a conchostracan fauna of the *Eosolimnadiopsis gallegoi* Zone (Fig. 5), which so far is unknown outside the Germanic Basin.

The richest conchostracan faunas of the Weser Formation, and especially of its marginal correlatives, occur in the Lehrberg Beds and correlatives (Fig. 5). During this time a distinct marine ingression took place, and dolomites and evaporite-bearing marls were deposited in the southern and central Germanic Basin. These beds record brackish deposition toward the basin margin, and freshwater lake deposition in some marginal areas, as well as in the northeastern Germanic Basin. The brackish and freshwater beds are rich in conchostracans, especially *Laxitextella seegisi* Kozur, the index species of the *L. seegisi* Zone (Fig. 5). The Lehrberg beds also contain important vertebrate fossil faunas, especially at the famous locality Krasiejów (Opole Silesia, southwestern Poland). These vertebrates (Dzik and Sulej, 2007) can be assigned to the early Adamanian land-vertebrate faunachron (Kozur and Weems, 2007, 2010a; Lucas, 2010c). The Krasiejów locality is especially important because *L. seegisi* and other conchostracans are present there (Olempska, 2004; Kozur and Weems, 2007). In the hypersaline marine and brackish Lehrberg Beds of the central and southern Germanic Basin, *Simeonella nostoria* Monostori, a euryhaline ostracod guide form of the lower Tuvalian of Hungary (Monostori, 1994) and the Alps also occurs. It was first discovered in southern Germany and well documented by Seegis (1997), but at that time it was still assigned to the upper Julian form *Simeonella alpina* Bunza and Kozur (see Kozur and Weems, 2007).

L. seegisi is also common in the USA. It is present in the *Anyuanestheria wingatella* Zone in the Bluewater Creek Formation at the Lake Ciniza locality in New Mexico and occurs also in the *Howellisaura princetonensis* Zone of the lower Cumnock Formation in the Sanford Subbasin of the Deep River Basin (Figs. 6 and 7). The *H. princetonensis* Zone occurs also in the lower fourth of the Lockatong Formation (up to the top of the Princeton Member) in the Newark Basin. There it was correlated with the middle "Lacian" (middle part of lower Norian) by palaeomagnetic correlation (e.g. Krystyn et al., 2002; Gallet et al., 2003; Muttoni et al., 2004). However, the combined correlation with euryhaline marine ostracods and conchostracans show that this part of the Adamanian belongs to the upper part of the lower Tuvalian (Kozur and Weems, 2007, 2010a).

The next younger diverse conchostracan fauna of the Germanic Basin is from the Coburg Sandstein in the Hassberge Fomation (Fig. 5). The index species of the *L. freybergi* Zone, *Laxitextella dorsorecta* (Reible), *Euestheria hausmanni* (Schmidt) and *L. freybergi* Kelber and Kozur (in Kozur and Weems, 2007), were described from this fauna. This is a typical late Carnian conchostracan fauna with the Carnian

index genus *Laxitextella*. The advanced *Laxitextella freybergi* distinguishes it from the lower Tuvalian faunas with *Laxitextella* of the *L. laxitexta* group, but the differences from the lower Norian faunas are even more pronounced. In the USA, *L. freybergi* is also known from the basal Gettysburg Formation of the Gettysburg Basin, which correlates to the upper Lockatong Formation (Smith Corner Member). This level has been correlated with the uppermost lower Norian by palaeomagnetic correlations (e.g., Krystyn et al., 2002; Gallet et al., 2003; Muttoni et al., 2004), but in the Germanic Basin this fauna lies immediately below the beginning of the late Tuvalian sporomorph association with *Classopollis meyerianus* (Klaus) de Jersey, Zhang & Grant-Mackie and *G. rudis* Venkatachala & Góczán in the overlying Mainhardt Formation and correlatives (Orłowska-Zwolińska, 1983; Heunisch, 2005; Schulz and Heunisch, 2005) (Fig. 5).

The youngest Carnian conchostracan fauna of the Germanic Basin is from the Heldburg Gypsum and correlatives (Kozur and Weems, 2010a). It is nearly identical with the conchostracans from the *L. freybergi* Zone of the Coburg Sandstein, but *Palaeolimnadia* n. sp. is also present, which occurs in monospecific assemblages in the basal Arnstadt Formation (basal Norian), and in the USA, in the Warford Member of the lower Passaic Formation (basal Norian as well). *Palaeolimnadia* n. sp. was assigned to *P. schwanbergensis* Reible by Kozur and Weems (2010a), but restudy of the material has shown that only deformed specimens are similar in outline to *P. schwanbergensis*, whereas undeformed specimens have an outline similar to *Shipingia* and are a new species that will be described by Kozur and Weems elsewhere.

The basal Arnstadt Formation in the Germanic Basin and the Warford Member of the Passaic Formation in Newark Basin contain a monospecific conchostracan fauna of *Palaeolimnadia* n. sp. (Figs. 5 and 6). All typical Carnian taxa have disappeared in this fauna and do not recur in the succeeding lower Norian *Euestheria buravasi–Euestheria* n. sp. Zone (Kozur and Weems, 2010a). The assignment of the Norian base by conchostracans to the base of the Warford Member (lower Passaic Formation) in the Newark Basin coincides closely with the assignment of the Norian base by vertebrates (see below) and by sporomorph associations, i.e., the *C. secatus* HO and *C. veruccosus* FO (see above). In the Germanic Basin, the next younger conchostracan association of the lower, but not lowermost Arnstadt Formation and correlatives (lower Stubensandstein) belongs to the *E. buravasi–Euestheria* n. sp. Zone (Kozur and Weems, 2010a) (Fig. 5). It occurs also in the Triangle Brick Quarry, Durham Subbasin of the Deep River Basin (North Carolina). *E. buravasi* Kobayashi is also present in the lower Norian of Thailand (Kobayashi, 1954).

The conchostracan correlation of the Newark Supergroup with the Germanic Basin by Kozur and Weems (2007, 2010a) has shown that the previous vertebrate palaeontological/palynological correlation of the Norian base in the Newark Supergroup was correct. We note that in placing the Carnian–Norian boundary as shown in Fig. 5, Kozur and Bachmann (2010b) assumed an age of 222 Ma for the Norian base

using the lowest value for the error range of the 225 ± 3 Ma for the conodont-dated basal Norian by Gehrels et al. (1986, 1987).

4.4. Tetrapods

Triassic tetrapod (amphibian and reptile) fossils have long been used in nonmarine biostratigraphy. Lucas (1998, 1999; Lucas et al., 2007; Lucas, 2010c) presented a comprehensive global Triassic tetrapod biochronology that divided the Triassic into eight time intervals (land-vertebrate faunachrons: LVFs) based on tetrapod evolution. This biochronology is based on an exhaustive review of an extensive array of publications wherein the localities yielding specific taxa in question are placed in detailed stratigraphic context (e.g., Heckert, 2004; Heckert and Lucas, 1998; Lucas et al., 1998; Heckert and Lucas, 1999, 2002; Heckert et al., 2006, 2008; see also references of Lucas, 1998; Lucas and Heckert, 2000; Lucas et al., 2007). The tetrapod biochronology divides Late Triassic time into four LVFs (ascending): Otischalkian, Adamanian, Revueltian and Apachean (Figs. 5–8). Huber et al. (1993) introduced provincial LVFs for the Newark Late Triassic tetrapod record (ascending): Sanfordian (~Otischalkian), Conewagian (~Adamanian) and Neshanician–Cliftonian (~Revuletian–Apachean) (Fig. 5)

Recently, Irmis et al. (2010) have criticized the global applicability of the tetrapod biochronology advocated here based on their disagreements with Lucas and others concerning tetrapod taxonomy in combination with their acceptance of abiotic chronologic techniques such as magnetostratigraphy and radioisotopic chronology. Key to their fossil identifications are a reliance on phylogenies (not always published) to deduce specific character states of taxa, rather than a strict identification of taxa based on morphology. For example, Lucas (1998) and Lucas and Heckert (2000) follow Hunt and Lucas (1991) in identifying a phytosaur with nares anterior to the antorbital fenestra as *Parasuchus* (= *Paleorhinus*) as this is the only taxon in which this character is known to occur, regardless of inferred phylogenies of phytosaurs. In contrast, Irmis et al. (2010) advocate basing the taxonomy of biostratigraphically-useful tetrapod taxa on the polarity of character states in a cladistic phylogeny. Here, we continue to use a tetrapod biostratigraphy/biochronology based on identifying tetrapod fossils by their morphology, not by inferred phylogenetic relationships.

Correlation of nonmarine tetrapod fossils to the standard global chronostratigraphic scale indicates that the Otischalkian and Adamanian LVFs are of Carnian age, and the Revueltian and Apachean LVFS encompass the Norian–Rhaetian (Fig. 8). Correlation of the LVFs to the Carnian–Norian boundary is aided by records of nonmarine tetrapod taxa in marine strata (Lucas and Heckert, 2000; Lucas, 2010c). Thus, there are two records of Otischalkian tetrapod index taxa in marine strata in Austria that support an Otischalkian–Carnian correlation. Koken (1913) described the temnospondyl amphibian *Metoposaurus santaecrucis* from a conglomeratic sandstone in the upper part of the Raibler Formation of Austria (also see Hunt, 1993). This is a Julian record, and thus correlates part of the Otischalkian (index taxon = *Metoposaurus*) to the Julian. Also from Austria, Huene (1939) described a skull fragment of the phytosaur *Parasuchus* (= *Francosuchus*) from the lower part of the Opponitzer Formation (Kalk) near Lunz (Hunt and Lucas, 1991). The occurrence is of late Carnian (Tuvalian) age, but it cannot be tied precisely to a particular ammonite zone.

Two Italian records are critical to correlation of the Revueltian to part of the Norian. One record is from the Zorzino Limestone (Calcare de Zorzino), which has been correlated to the mid-Norian (uppermost Alaunian) *Himavatites columbianus* ammonite zone (Jadoul et al., 1994; Roghi et al., 1995). Nonmarine tetrapods from this unit at the Cêne and Endenna quarries in Lombardy include the Revueltian index taxa *Mystriosuchus*, *Aetosaurus*, and *Eudimorphodon* (Wild, 1989; Lucas et al., 1998; Renesto, 2006). The other record is from the Forni Dolomite (Dolomia di Forni) in northeastern Italy, which is the same age as the Zorzino Limestone, i.e. mid-Norian (Roghi et

al., 1995). Its nonmarine tetrapods include the Revueltian index taxon *Eudimorphodon* (Dalla Vecchia, 1994). The Italian records thus provide direct evidence that at least part of the Revueltian = middle Norian (Alaunian).

In contrast, no such marine records are known for Adamanian tetrapod fossils. Their late Carnian age has been advocated in large part by their co-occurrence with late Carnian palynomorphs and conchostracans in the Chinle, Newark and Keuper sections, as described above (e.g., Lucas, 1998, 1999; Kozur and Weems, 2005, 2007; Lucas, 2010c; Kozur and Weems, 2010a). Assigning the Adamanian tetrapods a late Carnian age also has long been consistent with age assignments based on magnetostratigraphy (i.e. the Chinle–Newark correlation prior to 2002) and sequence stratigraphy (see reviews by Lupe and Silberling, 1985; Lucas and Marzolf, 1993; Lucas and Huber, 1994; Lucas, 1997, 1998, 2010c). Most significantly, no tetrapods assigned to the Adamanian occur in strata assigned a Norian age based on other biostratigraphic markers. Rather, only Revueltian tetrapods occur in rocks of demonstrable Norian age (see above). Consequently, the Adamanian–Revueltian boundary has been correlated to the Carnian–Norian boundary since Lucas and Hunt (1993) introduced the terms Adamanian and Revueltian. Of course, there is no reason to believe *a priori* that the marine and nonmarine boundaries should be exactly coeval, but the best available evidence (e.g., cross-correlation based on marine occurrences of nonmarine taxa) indicates that the Adamanian is Carnian and the Revueltian is Norian.

With regard to the marine occurrences of nonmarine tetrapods documented by Lucas (1998) and Lucas and Heckert (2000), Irmis et al. (2010, p. 45) attempt to cast doubt on the resulting correlations by stating "there is absolutely no reason to assume that taxa were "afraid" or unable to cross chronostratigraphic boundaries" and go on to argue that the presence of a taxon in Tuvalian (Carnian) age rocks does not preclude its range from extending back to the Middle Triassic or extending upward into the Norian. While we agree that it is certainly possible that the nonmarine range of taxa will not exactly coincide with the entirety of a particular marine interval, correlations are based on the actual occurrence of taxa, not their possible undocumented ranges.

In the Chinle Group, the base of the Revueltian is approximately at the base of the Painted Desert Member of the Petrified Forest Formation (Fig. 7). This interval is correlative, on lithostratigraphic and biostratigraphic grounds, to the base of the Bull Canyon Formation in eastern New Mexico, which yields fossils of *Aetosaurus* stratigraphically low in the formation (Heckert and Lucas, 1998). In the Newark Basin, *Aetosaurus* has its FO in the Warford Member of the Passaic Formation, and this marks the local base of the Revueltian (Lucas et al., 1998) (Fig. 6). In the Germanic Basin, the FO of *Aetosaurus* is in the basal Löwenstein Formation (lower Stubensandstein), which thus defines this as the base of the Revueltian interval in that section (Lucas et al., 1998; Lucas, 1999, 2010c) (Fig. 5). Adamanian tetrapods are known in the Chinle, Newark and Keuper below the FOs of *Aetosaurus*, so the base of the Revueltian/base of the Norian is approximated by the FO of *Aetosaurus* (Fig. 8).

The level of the oldest Norian vertebrates, for example *Aetosaurus* in the Warford Member, is very close to the disappearance of Carnian palynotaxa, such as *B. striata* and *C. secatus*, and the appearance of Norian taxa (e.g. *C. verrucosus*) within the interval of the Warford Member and the slightly younger Graters Member of the lower Passaic Formation. The vertebrate-based Norian base in the Warford Member (FO of *Aetosaurus*) perfectly coincides with the base of the monospecific *Palaeolimnadia* n. sp. Zone (in the sense of the *Palaeolimnadia schwanbergensis* Zone of Kozur and Weems, 2010a, see above), which is also in the Warford Member and is the Norian base as established with conchostracans. This conchostracan-based Norian base occurs in the Germanic Basin in a thin interval of the basal Arnstadt Formation (basal Steinmergelkeuper). The basal Arnstadt Formation belongs to the basal Norian and it corresponds with the

lower part of the Stubensandstein 1, which includes the lowest occurrence of *Aetosaurus* in the Germanic Basin. *Palaeolimnadia* n. sp. begins in the Germanic Basin within the uppermost Carnian Heldburg Gypsum Member, but there it is accompanied by typical late Tuvalian conchostracans, such as *L. freybergi* Kelber and Kozur. Summarizing, we can say that the tetrapod biostratigraphy places the Carnian–Norian boundary in the same stratigraphic interval as the palynostratigraphy and conchostracan biostratigraphy (Fig. 8).

5. Cyclostratigraphy

Van Houten (1962, 1964) noted an apparent periodicity in the sandstone-mudstone-shale cycles of the Late Triassic Lockatong Formation of the Newark Basin and proposed that the sequence represented transgressive–regressive lacustrine cycles. Furthermore, Van Houten used an estimated an average sedimentation rate, derived from varve counts, to determine that the frequency of the cycles approximated the precessional signal of the Milankovitch orbital frequencies. Olsen (1984, 1986), Olsen et al. (1989) and Olsen and Kent (1996) further developed the idea that Milankovitch-frequency cyclicity is recorded by the lacustrine cycles in Newark Basin strata. Olsen tested this hypothesis in the transgressive–regressive cycles of the Newark Basin by dividing the cycles into seven lithologies and assigning a depth ranking to each lithology. By using Fourier analysis to examine the recurrence interval of the lithofacies and generating power spectra of the sedimentary cycles, peak recurrence intervals could be tested statistically. Further, calibrating the cycle thicknesses to the sedimentation rate employed by Van Houten permitted assignment of a period of 18–25 kyr to the basic precession cycle, termed the "Van Houten cycle" (Olsen et al., 1989). The power spectra also identified compound cycles originally identified by Van Houten and allowed assignment of these to the frequencies of the eccentricity cycles at 95 kyr and 125 kyr, the long eccentricity cycle at 400 kyr, and the very long cycle of ~2000 kyr.

The original work of Van Houten (1962, 1964) was limited to the Lockatong Formation, but Olsen extended the interpreted cyclostratigraphy to include all of the Triassic–Jurassic strata of the Newark Supergroup. The Newark Basin Coring Project produced nearly 7 km of stratigraphically-overlapping core from seven laterally offset drill holes. Olsen and Kent (1996) and Olsen et al. (1996) analyzed the ostensibly complete sections of the Lockatong and Passaic formations obtained from the project, applying the depth-rank analysis technique to conclude that the individual, 4–7 m-thick Van Houten cycles are the most obvious component of cyclicity at the outcrop scale, but statistically they are more weakly expressed than the thicker, modulated cycles. In the cored strata, Olsen and Kent (1996) and Olsen et al. (1996) defined: (1) short modulating cycles (~100 kyr), each comprising five Van Houten cycles, (2) intermediate modulating cycles (totaling about 400 kyr long), which they termed McLaughlin cycles, each comprising four short modulating cycles, and (3) long modulating cycles (about 2 myr long) of four to five McLaughlin cycles. Within the Upper Triassic section of the Newark Basin, they identified 60 McLaughlin cycles in the upper Stockton through Passaic formations that should therefore represent 24 myr of sedimentation (Kent and Olsen, 1999) (Fig. 4).

If the interpretations of the cycle durations are valid, the cyclostratigraphy of the Newark Basin (and by analogy much of the Newark Supergroup) is a determinative chronostratigraphy, providing what has been termed an astronomically-tuned timescale for the Late Triassic (Kent and Olsen, 1999). Given the radioisotopic ages of the Newark volcanics, which average ~201 Ma (cf. Marzoli et al., 2011), as a reference datum, and *assuming no hiatuses in the sedimentary succession*, cycle counting should yield an absolute age for the strata in these basins (Fig. 4). In theory, correlation of the thus-calibrated section to a biostratigraphically-calibrated section then would allow determination of the absolute ages of the stage boundaries. Such cycle counting identified the palynological Carnian-Norian boundary as being ~216–218 Ma (based on a long-accepted age of 202 Ma for the lower CAMP lava flow), and, as noted above, this became the basis for assigning this age to the Norian base for about a decade (Fig. 3). For the base of the Norian at the base of the Warford Member of the Passaic Formation, this strict counting would yield an estimated age of ~217.8 Ma (Fig. 4).

There are, of course, reasons to doubt the accuracy of this date based on the astronomically-tuned timescale. Foremost is the question of the completeness of the Newark Basin stratigraphy. In particular, Kozur and Weems (2005, 2007, 2010a) correlated the conchostracan biostratigraphy of the Newark Supergroup with the Germanic Triassic and concluded that the section is neither complete nor appropriately dated. They concluded that the uppermost Norian and most of the Rhaetian are missing in the Newark Basin and elsewhere in the Newark Supergroup, which may mean that about 3–4 myr of Late Triassic time is not represented by the Newark cycles. In that case, the cycle counting plus the inferred 3–4 myr hiatus would suggest a Carnian–Norian boundary age of 219–222 Ma.

Tanner (2010) also raised significant questions regarding the validity of the cyclostratigraphic model. Tanner (2010) noted that the calculations of cycle duration from varve counts relied on the estimate of the sedimentation rate measured from the thickness of the varves in the deepest lacustrine facies (Olsen, 1986) when the cycle actually comprises a variety of lithologies deposited in littoral and alluvial environments. Much of the Passaic Formation, for example, is dominated by alluvial deposits and lacks the distinctive lacustrine cycles that characterize the Lockatong Formation, making identification of cycles in this part of the stratigraphic section very subjective.

In particular, Tanner (2010) critiqued the assumption that sedimentation in the Newark Basins was controlled entirely by orbitally-forced climate change to the exclusion of other mechanisms such as tectonic control of base-levels. As noted above, the calculation of the Carnian–Norian boundary by counting McLaughlin cycles presumed that no significant unconformities or hiatuses occurred throughout the history of sedimentation in the basin over a span of at least 16 myr (see discussion in Tanner, 2010). Despite these inherent problems, we regard the method of cycle counting as a way to arrive at a first approximation of the numerical age of the base of the Norian. Assuming an age of 201.7 Ma for initiation of CAMP activity in eastern North America (Dunning and Hodych, 1990; Hames et al., 2000; Kent and Olsen, 2000; Nomade et al., 2007; Jourdan et al., 2009; Marzoli et al., 2011), a Rhaetian duration of 4 myr (e.g., Ogg, 2004) and an intra-Rhaetian gap in the Newark section of about 3.8 myr, counting 40 cycles below the first lava flow yields an age that is almost 4 million years older than the previous age of 217–218 Ma.

Kozur and Weems (2007, 2010a, 2010b) correlated the base of the Hettangian, which is defined in marine strata as the FAD of the first psiloceratid ammonite, *Psiloceras spelae* Guex, Taylor, Rakus & Bucher, to continental beds at the FAD of the conchostracan genus *Bulbilimnadia* Shen based on documentation of its oldest species at the base of the *B. killianorum* Zone (Kozur and Weems, 2010a, 2010b; Lucas et al., 2011). Based on this correlation, Kozur and Weems (2005) pointed out that the lowest CAMP lava flow of the Newark Supergroup is late Rhaetian in age (also see Lucas and Tanner, 2007; Cirilli et al., 2009).

The question of how much younger is the Hettangian base (equivalent to the FAD of *P. spelae*) than the base of the lowest lava flow is critical to downward counting of Milankovitch cycles in the Newark Supergroup. Whiteside et al. (2010) used $\delta^{13}C_{org}$ stratigraphy to identify the base of the Hettangian in the Newark Supergroup, which confirmed the placement of the Triassic-Jurassic boundary by conchostracan biostratigraphy (Kozur and Weems, 2007, 2010a). Whiteside et al. (2010) correlated this boundary with the Milankovitch cyclicity and stated that it lies about 100,000 years above the lowest lava flow.

Radioisotopic age data of Schaltegger et al. (2008) in Peru have shown that the Hettangian base is at ~201.5 Ma (Schoene et al., 2010 offer a slightly younger estimate of ~201.3 Ma). Therefore, using the estimate of Whiteside et al. (2010), the age of the top of the first lava flow is ~201.6 Ma. If the eruption of the lowest lava flow did not occupy a time interval greater than 100,000 years, the base of CAMP volcanism in the Newark Supergroup is not older than 201.7 Ma. Kozur and Weems (2005, 2007, 2010a) have demonstrated previously that the horizon below the lowest CAMP flow formerly considered the position of the base of the Jurassic, which is about one precession cycle (~20,000 years) below the lowest lava flow (e.g., Olsen and Kent, 1996), coincides with a gap that separates late Sevatian conchostracan faunas below from (late) Rhaetian conchostracan faunas above (Fig. 6). This means that the preserved portion of the Rhaetian in the Newark Supergroup has a duration of only about 200,000 years. This conclusion is supported by the fact that all palaeontologically-proven Rhaetian beds in the Newark Supergroup have a normal magnetic polarity, whereas at least a third of the entire marine Rhaetian is reversed (Muttoni et al., 2010; Hüsing et al., 2011).

In the Newark, the very short reversed interval immediately below the Hettangian base was not sampled for magnetostratigraphy, as stated by Whiteside et al. (2010). Except for this very short reversed interval, the entire upper part of the upper Rhaetian is of normal polarity (Fig. 4), so the magnetostratigraphy of the Newark Supergroup indicates that at least part of the upper Rhaetian is present. If we assume that the Rhaetian has a duration of ~4 myr (cf. Ogg et al., 2008), then as much as 3.8 myr of the Rhaetian is missing in the Newark section, much of it at the gap at the formerly inferred Triassic-Jurassic boundary (at the "Passaic palynofloral event" in Fig. 6) and possibly another gap is present below the first lava flow. This 3.8 myr thus has to be added to the counting down from 201.7 Ma and this would give a Norian base (at the base of the Warford Member) of 217.5 + 3.8 Ma, or 221.3 Ma.

There are two possible problems with this counting. The first is, as discussed above, whether or not the Milankovitch cyclicity is reliable and whether there are no gaps in the sequence. Where we have checked the data, the Milankovitch cycles used in Kent and Olsen (2000) appear to be reliable. Kent and Olsen (2000) have indicated that the base of the Newark Supergroup is at 233 Ma. If we count the cycles down from 201.7 Ma, this would give an age of 232.7 Ma. Adding the Rhaetian gap of about 3.8 million years then yields an age of 236.5 Ma. Significantly, Kozur and Weems (2007, 2010a, 2010b) have shown biostratigraphically that the base of the Newark Supergroup coincides with the base of the Carnian, which lies around 237 Ma (Bachmann and Kozur, 2004). Without applying the Rhaetian gap, the base of the Newark would be much younger (232.7 Ma) than the base of the Carnian, but with the Rhaetian gap, the base is only slightly younger. However, Kozur and Weems (2007, 2010a) also indicate a widespread gap above the "middle" Carnian (Cordevolian) in the Newark section (Fig. 6), which, if compensated for, pushes the base of the Newark closer to 237 Ma. Thus, seemingly, the preserved Milankovitch cycles are reasonably calibrated and, except for the long gap in the Rhaetian and the shorter gap after the Cordevolian, probably no substantial gaps are present in the long Carnian to Rhaetian interval of the Newark Supergroup.

The second problem with using cycle counting is the duration of the Rhaetian, because very different Rhaetian durations have been proposed previously. Krystyn et al. (2002) and Gallet et al. (2003) assumed only 2 myr for the Rhaetian, while Kozur (2003) assumed 4.4 myr, Gradstein et al. (2005) and Ogg et al. (2008) used a length of 4 myr, Bachmann and Kozur (2004) 5.4 myr, Muttoni et al. (2004) 8 myr and Hüsing et al. (2011) assumed a duration of 9 myr. The 2 myr duration of the Rhaetian, which is apparently too short, resulted in part from the fact that the Rhaetian base used in Krystyn et al. (2002) and Gallet et al. (2003) was too high, and caused a considerable part of the lower

Rhaetian (the entire *Paraochloceras suessi* Zone) to be assigned to the so-called Sevatian 2. However, Krystyn and Kuerschner (2005) adopted a Rhaetian base with the FAD of *Misikella posthernsteini* at the base of the *P. suessi* Zone, as first established by Kozur (1996). Just by this correction of the Rhaetian base, the duration of the Rhaetian would change from 2 myr, as advocated by Krystyn et al. (2002) and Gallet et al. (2003), to 3–4 myr. This is the basis for a duration of the Rhaetian of around 4 myr assumed by most authors, which we follow for the cycle counting in the present paper. A duration of 8 (or 9) myr for the Rhaetian is seemingly too long because if we apply this duration and include the 200,000 years of fossil-proven Rhaetian in the Newark Supergroup, we add 7.8 million years to the interval from the base of the CAMP volcanism (201.7 Ma) to the base of the Newark Supergroup (232.7 Ma), which would then become 240.5 Ma (= base of the Carnian); this would be close to the age of the Anisian–Ladinian boundary (Mundil et al., 2010). Therefore, a duration of 8 myr for the Rhaetian seems unacceptable.

6. Magnetostratigraphic correlation

The global polarity timescale for rocks of Late Jurassic, Cretaceous and Cenozoic age provides a valuable tool for evaluating and refining correlations that are based primarily on radioisotopic ages or biostratigraphy. However, there is no agreed geomagnetic polarity timescale (GPTS) for the Triassic, although a composite GPTS is now becoming available based on successions assembled from marine and nonmarine sections in North America, Europe, and Asia (see extensive recent review by Hounslow and Muttoni, 2010).

Magnetostratigraphic studies across the Carnian–Norian boundary in marine strata began in the early 1990s, and data are now available from the Bolücektasi Tepe and Kavaalani sections in Turkey (Gallet et al., 1992, 1993, 2000), the Pizzo Mondello section in Sicily (Muttoni et al., 2001, 2004), the Silická Brezová section in Slovakia (Channell et al., 2003) and the Guri Zi section in northern Albania (Muttoni et al., 2005). These Carnian–Norian boundary sections are all relatively thin, with a total thickness of Norian strata of less (and usually much less) than 300 m, in limestone-dominated sections in which biostratigraphic control is primarily based on conodonts (e.g., Fig. 9). The magnetostratigraphy of these sections documents that the Norian base (based on current biostratigraphic concepts, see above) is consistently in an interval of normal polarity. The challenge has been to correlate them to the magnetostratigraphy from the much thicker Newark Supergroup section (total Norian thickness at least 2 km) in which precise independent (noncyclostratigraphic) age control is much less available.

None of the published magnetostratigraphy across the Carnian–Norian boundary in these marine sections can be unambiguously correlated to the nonmarine Newark magnetostratigraphy. In part, this is because there is no unique pattern to the magnetic polarity history across the Carnian–Norian boundary that can be recognized independent of the effects of sedimentation rates. Given that average sedimentation rates in the Newark typically must be at least an order of magnitude greater than those of the correlative marine sections, it is hardly surprising that a unique pattern match has not been found based on magnetostratigraphy alone. Furthermore, the presence/absence of unconformities (hiatuses) in the Newark (see above) and the marine sections has not been fully investigated, so that the completeness of the polarity history recorded by the various sections remains open to question. Indeed, the fact that the marine sections contain far fewer magnetochrons than does the presumed age-equivalent interval of the Newark (Fig. 9) is *prima facie* evidence that they are incomplete records, despite their apparent biostratigraphic "completeness." Finally, disagreements over establishing biostratigraphic tiepoints between the Newark and the marine sections have further hindered their magnetostratigraphic correlation by eliminating agreement on the
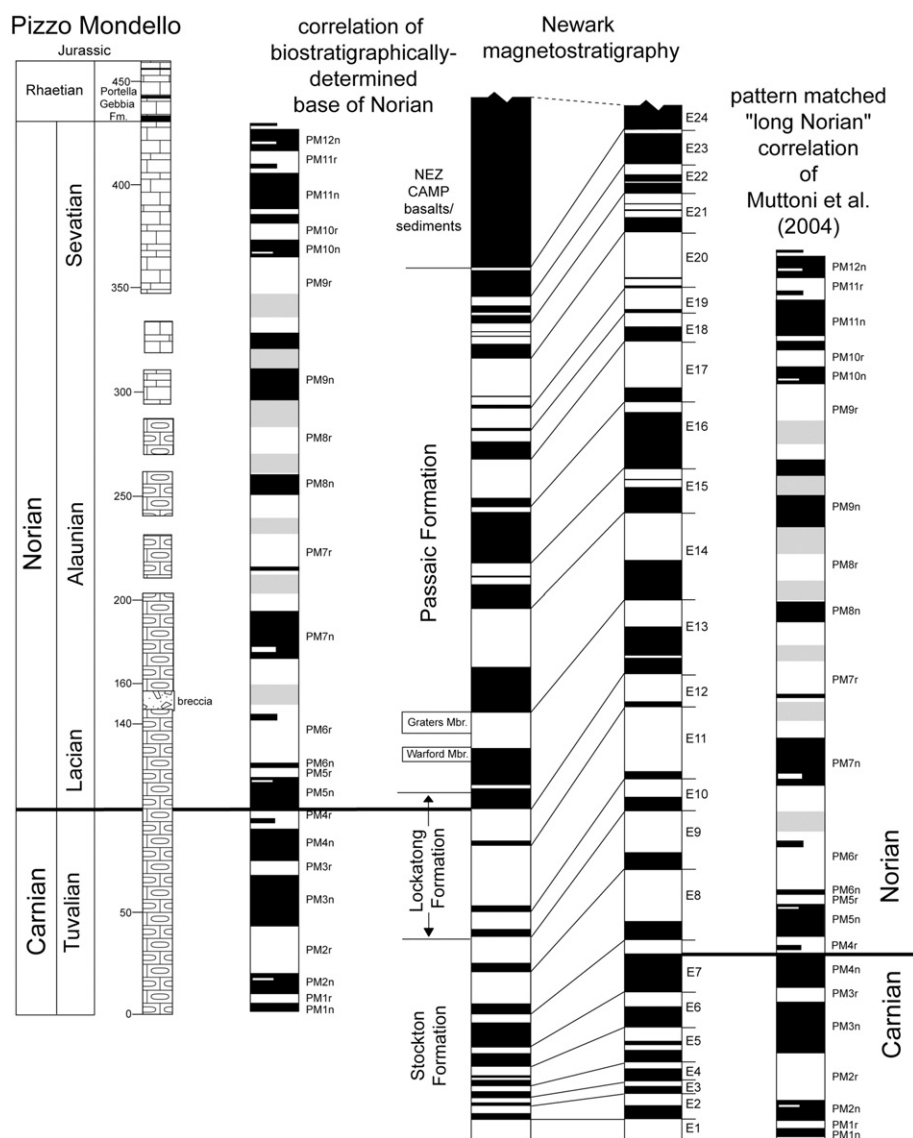
Fig. 9. Magnetostratigraphic correlations of the Pizzo Mondello and Newark sections. On the left, the correlation matches the marine and nonmarine, biostratigraphically-determined Carnian–Norian boundary. On the right is the "pattern matched" correlation of Muttoni et al. (2004), which became the basis of the "long Norian".

biostratigraphic datum points that are often essential to robust correlations of magnetostratigraphy.

Prior to Krystyn et al. (2002), Channell et al. (2003) and Muttoni et al. (2004), the palynostratigraphic placement of the Carnian–Norian boundary near the base of the Passaic Formation in the Newark Supergroup was used as a biostratigraphic datum upon which to base magnetostratigraphic correlations. This located the Carnian–Norian boundary within Newark magnetic polarity chron E13, an interval of relatively long normal polarity followed by reversed polarity (Fig. 4). Nevertheless, attempts to correlate this polarity chron to the marine sections, well reviewed by Krystyn et al. (2002) and Channell et al. (2003), were fraught with problems simply because no convincing match of the polarity patterns could be achieved, even with the help of a biostratigraphic datum. Thus, various correlations were proposed, but no consensus was reached on correlation of Newark magnetostratigraphy to the magnetostratigraphy of coeval marine sections (Krystyn et al., 2002; Channell et al., 2003).

Krystyn et al. (2002) and Muttoni et al. (2004) took a very different approach by arbitrarily abandoning the palynostratigraphic placement of the Carnian–Norian boundary near the base of the Passaic Formation in the Newark Supergroup and correlating strata solely by attempting to match polarity patterns (Fig. 9), an approach

reiterated by Hüsing et al. (2011). Their preferred correlation led Muttoni et al. (2004) to place the base of the Norian stratigraphically much lower in the Newark column than had ever been done before, close to the base of the Lockatong Formation with an estimated age of about 228 Ma based on counting McLaughlin cycles (Fig. 4). Krystyn et al. (2002) and Gallet et al. (2003, 2007) placed the Norian base at the base of the Adamanian, low in the Stockton Formation with an estimated age of 228 Ma. Channell et al. (2003) and Muttoni et al. (2004) have chosen a Norian base higher in the Adamanian, in the upper Stockton Formation close to the base of the Lockatong Formation. Channell et al. (2003) estimated the age of the Norian base at 226 Ma, while Gallet et al. (2003) estimated it at 227 Ma and Gallet et al. (2007) estimated it at 229 Ma.

Earlier, Muttoni et al. (2001) presented a magnetostratigraphy of the Carnian–Norian boundary at Pizzo Mondello based on sampling only part of the exposed section and with relatively imprecise biostratigraphic control of the position of the Carnian–Norian boundary. They correlated these data to the Newark section based on visual matching of magnetozones, ignoring the palynologically-placed Carnian–Norian boundary in the Newark. Thus, Muttoni et al. (2001, Fig. 7) correlated an ~120 m thick section at Pizzo Mondello to a ~1.5 km thick Newark section to arrive at a correlation that placed the Norian base at the

boundary of Newark magnetozones E15/E16, which is in the middle of the Passaic Formation (Fig. 9).

Subsequently, Muttoni et al. (2004) sampled much more of the Pizzo Mondello section (~450 m). They also had more precise biostratigraphic control of the position of the Carnian–Norian boundary (Fig. 9), which is approximated by the LO of *Metapolygnathus communisti*, considered by some as a proxy for the ammonoid-based boundary (Krystyn et al., 2002). Although this provided a longer magnetic polarity record at Pizzo Mondello to correlate to the Newark, it also made it more difficult to arrive at a correlation by simple visual matching of magnetozones. As they had done in 2001, Muttoni et al. (2004) ignored the palynologically-based Carnian–Norian boundary in the Newark as a datum upon which to correlate the magnetostratigraphy. Thus, they stated that "….palynological data for the New Oxford–Lockatong palynofloral zone are in fact compatible with either a late Carnian or Norian age and lack definitive Carnian taxa such as *C. secatus* (e.g., Litwin and Skogg, 1991)," despite the fact that all palynologists have long regarded the New Oxford–Lockatong palynofloral zone as late Carnian (see especially Cornet and Olsen, 1985; contra Irmis et al., 2010).

Muttoni et al. (2004) correlated the Pizzo Mondello and Newark magnetostratigraphies by assuming that thickness in both sections is a proxy of time, which requires the simplifying (but indefensible) assumption that sedimentation rates in both sections remained constant throughout their histories. They then evaluated 16 possible correlation matches of the shorter Pizzo Mondello polarity record with the Newark record. For each match, they calculated a linear correlation coefficient that related the thicknesses of each of the 26 zones at Pizzo Mondello to the Newark zones, and a t value was calculated to evaluate the quality of each correlation (see Lowrie and Alvarez, 1984 for this method). The analysis identified two options as having the highest t values: (1) a "long Carnian," which is the same correlation advocated by Muttoni et al. (2001), namely that the Carnian–Norian boundary is at the boundary of the E15/E16 magnetozones in the Newark; and (2) a "long Norian," in which the Carnian–Norian boundary correlated to approximately the top of Newark magnetozone E7 (Fig. 9). Given that the long Norian option had a higher t value (2.17) than did the long Carnian option (t = 2.0), the long Norian option was chosen by Muttoni et al. (2004).

If we honor the biostratigraphic position of the Carnian–Norian boundary in the Newark section, and use it as a biostratigraphic datum to correlate the magnetostratigraphy between Pizzo Mondello and the Newark, we arrive at quite a different correlation than those advocated by Muttoni et al. (2001, 2004) (Fig. 9). This correlation, however, faces the problem that at Pizzo Mondello most of the upper Tuvalian section is of normal polarity, whereas in the Newark Basin the section immediately below the palynologically-placed Carnian–Norian boundary is mostly of reversed polarity (Fig. 9). Of course, we do well to recognize that the mostly normal Tuvalian interval at Pizzo Mondello is ~100 m of cherty, and therefore diagenetically altered, limestone, whereas the apparently coeval interval of the Newark is more than 1 km of clastic redbed sediments.

There is no basis for believing that the condensed section at Pizzo Mondello is more complete than the Newark Basin section. Indeed, little discussed is the "breccia" interval in the lower Norian of the Pizzo Mondello section (Fig. 9). The significance of this unconformity-bounded unit and the potential for other disconformities within the section has not been explored. Consequently, we consider it illogical to expect that the ~400 m of limestone at Pizzo Mondello would produce a magnetic polarity record with the same relative thicknesses of magnetozones as the 3.5 km of clastic sediments (mostly mudrocks) in the Newark Basin; to assume this produces correlations that contradict defensible biostratigraphic correlations, as is clear from Muttoni et al. (2001, 2004). It has long been a basic principle of magnetostratigraphic correlation that a datum (biostratigraphic or radioisotopic) is needed to correlate magnetostratigraphic sections reliably, simply because magnetozone thicknesses vary greatly

with sedimentation rates, which typically confound a simple matching of magnetozones. Therefore, it is not surprising that the Pizzo Mondello section yields a magnetostratigraphic record that does not precisely or directly correspond to the Newark section either in reversal frequency or in pattern (Fig. 9).

Clearly, simple pattern matching of the magnetostratigraphy cannot produce a defensible correlation. Indeed, simple pattern matching of Tethyan marine magnetostratigraphy to the Newark magnetostratigraphy has always produced correlations that do not honor the biostratigraphically-determined position of the Carnian-Norian boundary in the Newark section (e.g., Krystyn et al., 2002, Fig. 4; Gallet et al., 2003, Fig. 1; Channell et al., 2003, Fig. 14; Hounslow and Muttoni, 2010, Fig. 12). Any correlation of the magnetic stratigraphy at Pizzo Mondello and the Newark Basin that can be strongly supported must equate the position of the Carnian–Norian boundary in each section, and this then equates Pizzo Mondello zone PM5 with Newark zone E13 (Fig. 9). This does not produce a convincing match of magnetozone thicknesses, undoubtedly because of vastly different sedimentation rates, but it does produce a plausible pattern match that, more importantly, is consistent with biostratigraphic constraints (Fig. 9).

Magnetostratigraphy of the Carnian–Norian boundary in the Chinle Group has been correlated to the Newark magnetostratigraphy by correlating the Adamanian–Revueltian boundary to the base of the Norian at the base of Newark magnetozone E13 (e.g., Molina-Garza et al., 1996; Steiner and Lucas, 2000). Although the pattern matching of these correlations is not definite, the Chinle sections do resemble the Newark (in this correlation) by being dominantly normal polarity above the Norian base, and mostly of reversed polarity below. Zeigler et al. (2008) recently reported a wholly reversed polarity interval across the Adamanian–Revueltian boundary in a Chinle section, but this is very different from the polarity pattern of other studies and may be a spurious result based on incomplete sampling (J. Geissman, pers. comm., 2010).

7. Radioisotopic ages

As mentioned above, there are very few reliable and precise radioisotopic ages that can be unambiguously related to the biostratigraphic position of the Carnian–Norian boundary. This situation has begun to improve recently, but presently available numbers (Table 1) still do not fix the numerical age of the boundary with certainty. This is clearly reflected by the recent review of the Triassic numerical timescale by Mundil et al. (2010), who concluded that the only certain conclusion from the radioisotopic ages is that the base of the Norian is younger than 230 Ma.

A series of recently published numerical ages are relevant to the age of the Carnian–Norian boundary; these are based on U/Pb dating of detrital zircons in the Chinle Group of the American Southwest by Riggs et al. (2003), Irmis and Mundil (2008), Dickinson and Gehrels (2008, 2009), Heckert et al. (2009), Ramezani et al. (2009, 2011) and Irmis et al. (2011) (Table 1). Except for Dickinson and Gehrels (2009), these articles indicate ages of about 220 Ma for strata in the upper part of the Adamanian interval, which is below the HO of *C. secatus* and thus biostratigraphically of late Carnian age. Stratigraphically higher zircons, from the Revueltian interval, are about 211–213 Ma in age, and these must be early or middle Norian ages based on the biostratigraphy.

Of considerable importance is that most radioisotopic dates from Chinle Group strata were obtained on grains isolated from fluvial sandstones rather than from ash beds; i.e., these zircons were, by definition, reworked, so the ages obtained should be regarded as maximum ages. Riggs et al. (2003), for example, obtained a U/Pb age of 213 ± 1.7 Ma for zircons from the Norian Black Forest Bed in the Painted Desert Member of the Petrified Forest Formation, but speculated that the actual date of deposition probably was closer to 209 Ma. The study by Ramezani et al. (2009, 2011) obtained zircon

Table 1
List of radioisotopic ages relevant to the numerical age of the base of the Norian.

Locality	Age	Method	Biostratigraphy	Reference
American Southwest (upper Chinle Group)	203–216 Ma[a]	U/Pb detrital zircons	Norian: tetrapods; pollen	Dickinson and Gehrels (2008)
Eastern USA (New Haven Arkose)	211.9 ± 2 Ma	U–Pb calcrete	Norian: tetrapods; pollen	Wang et al. (1998)
Arizona, USA (Black Forest Bed)	211.1 ± 0.7 Ma	U/Pb detrital zircons	Norian: tetrapods; pollen	Heckert et al. (2009)
Arizona, USA (Black Forest Bed)	213 ± 1.7 Ma	U/Pb detrital zircons	Norian: tetrapods; pollen	Riggs et al. (2003)
Argentina (Ischigualasto Fm.)	217.0 ± 1.7 Ma (~220 Ma)[b]	^{40}Ar/^{39}Ar	Carnian: tetrapods	Shipman (2004), Currie et al. (2009)
Arizona, USA (Black Forest Bed of Petrified Forest Member)	209.926 ± 0.072/0.13/0.26 Ma	U/Pb detrital zircons	Norian: tetrapods; pollen	Ramezani et al. (2011)
Arizona, USA (basal Petrified Forest Member)	213.124 ± 0.069/0.14/0.27 Ma	U/Pb detrital zircons	Norian: tetrapods; pollen	Ramezani et al. (2011)
Arizona, USA (upper Sonsela Member)	218.017 ± 0.088/0.15/0.28 Ma	U/Pb detrital zircons	Norian: tetrapods	Ramezani et al. (2011)
Arizona, USA (lower Sonsela Member)	219.317 ± 0.080/0.14/0.27 Ma	U/Pb detrital zircons	Norian: tetrapods	Ramezani et al. (2011)
New Mexico, USA (Blue Mesa Member)	218.1 ± 0.7 Ma	U/Pb detrital zircons	Late Tuvalian: tetrapods; pollen	Irmis et al. (2011)
New Mexico, USA (Blue Mesa Member)	219.7 Ma	U/Pb detrital zircons	Late Tuvalian: tetrapods; pollen	Irmis and Mundil (2008)
New Mexico, USA (Blue Mesa Member)	220.9 ± 0.6 Ma	U/Pb detrital zircons	Late Tuvalian: tetrapods; pollen	Heckert et al. (2009)
Arizona, USA (Blue Mesa Member)	223.036 ± 0.059/0.13/0.27	U/Pb detrital zircons	Tuvalian: tetrapods; pollen	Ramezani et al. (2011)
Arizona, USA (uppermost Mesa Redondo Formation)	225.185 ± 0.079/0.15/0.28	U/Pb detrital zircons	Tuvalian: tetrapods	Ramezani et al. (2011)
American Southwest (lower Chinle Group)	222–234 Ma[a]	U/Pb detrital zircons	Carnian: tetrapods;	Dickinson and Gehrels (2008)
West Texas, USA (upper part of Bull Canyon Formation)	~225 Ma	Rb–Sr on detrital mica	Norian: tetrapods, pollen	Long and Lehman (1993)
SE Alaska (Puppets Formation)	225 ± 3 Ma	U/Pb zircons in rhyolite	Latest Carnian-earliest Norian: conodonts	Gehrels et al. (1987)
Argentina (Ischigualasto Fm.)	227.8 ± 0.3 Ma (~230 Ma)[b]	^{40}Ar/^{39}Ar	Tuvalian: tetrapods	Rogers et al. (1993)
Italy (Pignola 2)	230.9 ± 0.33 Ma	^{206}Pb/^{238}U	Lower Tuvalian: conodonts	Furin et al. (2006)

[a] After eliminating contradictory or implausible ages (see text).
[b] With 1% correction added.

ages of ~219–223 Ma from the lowermost Norian Sonsela Member and the underlying upper Carnian Blue Mesa Member. These results are consistent with the results of similar studies by Irmis and Mundil (2008) and Heckert et al. (2009). On face value, these ages together indicate a Carnian–Norian boundary younger than 223 Ma, and older than 219 Ma.

Dickinson and Gehrels (2008) conducted an ambitious study of detrital zircon ages from the Chinle Group and equivalent strata, comprising 1808 individual analyses. The results of the study yield provenance information that provides important data on Late Triassic drainage patterns on a continental scale. Although the purpose of the original study was not related primarily to refining the Triassic time scale, the results do have bearing on the question of Triassic stage boundaries. While the bulk of the data includes ages of zircons derived from lower Palaeozoic to Precambrian continental sources, it also includes zircon ages from the volcanic arc that was active during Chinle Group deposition.

Dickinson and Gehrels (2009) thus incorporated the ages of arc-derived grains (from their 2008 study) into a broader study of the utility of depositional zircon ages for constraining formation ages. They present youngest single grain ages and mean age clusters (calculated at both σ and 2σ overlap) based on eight upper Chinle samples (Sonsela Member and correlatives, plus some younger strata) and 19 lower Chinle samples (Shinarump Formation and correlatives). For each sample, they analyzed approximately 100 detrital zircon grains. Each age they report has a 1σ error bar of at least 2 million years, and usually more, because they relied on laser-ablation, inductively-coupled plasma-mass-spectrometry (LA-ICP-MS) technique, which is adequate for most detrital zircon studies, but subject to greater discordance due to Pb loss than the chemical abrasion (CA-TIMS) technique.

Dickinson and Gehrels (2009, Table 1) upper Chinle samples (which are biostratigraphically Norian) have a youngest single grain age range of 203–254 Ma. The grain cluster means at 1σ yield values ranging from 209 ± 4 to 308 ± 5 Ma. If we discard the pre-Triassic ages (four samples), then this range is 209–225 Ma. In particular, the Sonsela Sandstone, which is correlated as lowermost Norian, yielded a cluster mean (from six grains) of 212 ± 2 at 1σ, and a mean of 216 ± 1 (based on 16 grains) at 2σ. The lower Chinle Group samples (these are biostratigraphically Carnian) have an age range of 215 ± 7 to 428 ± 2 Ma, and if we discard the implausibly old ages (242, 268, 309, 311 and 428 Ma), then the range becomes 215–234 Ma. According to Dickinson and Gehrels (2009), these ages are consistent with the "long Norian," and they place the base of the Norian at 227 Ma (Dickinson and Gehrels, 2009, Fig. 3).

However, we question this conclusion because of the wide scatter and temporal overlap of biostratigraphically Carnian and Norian ages in their samples. Thus, their biostratigraphically Norian detrital zircons (minus the discarded outlier) have a maximum age of 223 Ma, whereas their youngest Carnian detrital zircon age is 215 Ma, and this age is from a sample stratigraphically below their Norian samples. We can eliminate this contradictory overlap by discarding the oldest Norian age of 223 Ma, which reduces the range of Norian ages to 203–216 Ma, and discarding the youngest Carnian age, which reduces the Carnian age range to 222–234 Ma. If we accept these non-contradictory age ranges, then the Carnian–Norian boundary based on the data of Dickinson and Gehrels (2009) must be younger than 222 Ma; i.e., their data do not support a long Norian.

Key to all of this discussion is the presence of a robust Chinle lithostratigraphy (Stewart et al., 1972; Lucas, 1993) that, combined with an understanding of the biostratigraphy built upon it and described here, allows discrimination of plausible versus implausible dates. At present there is no way to determine a priori whether a detrital zircon will yield a reliable, essentially synsedimentary age, an older age reflecting a non-volcanic provenance, or a spurious, erroneously young age.

Although we accept that the detrital zircon method of Dickinson and Gehrels (2008, 2009) is inherently an imprecise method of determining the ages of deposition of sedimentary rocks, we note that their results are broadly consistent with other studies (Riggs et al., 2003; Irmis and Mundil, 2008; Heckert et al., 2009; Ramezani et al., 2009, 2011; Irmis et al., 2011) that seem to suggest a Carnian-

Norian boundary close to 220 Ma. The wide scatter and contradictory overlaps in the youngest detrital zircon age data of Dickinson and Gehrels (2009) may reflect problems of undetected lead loss and diagnetic alteration.

Indeed, Dickinson and Gehrels (2009) study provides a further caution in the use of detrital zircon ages because of the age of 230 Ma they report for a single sample from the Holbrook Member of the Moenkopi Formation in Arizona. This age is comparable to a U–Pb age of 233 Ma reported by Hunt and Lohrengel (2005) from the correlative upper red member of the Moenkopi Formation in Utah. Yet, as Dickinson and Gehrels (2009) acknowledge, the age of the Holbrook Member is well established by biostratigraphy and magnetostratigraphy as early Anisian (Steiner et al., 1993; Lucas and Schoch, 2002; Hounslow and Muttoni, 2010). This means that the Holbrook Member should be older than 240 Ma, based on current calibration of the numerical age of the Anisian (Mundil et al., 2010). As Dickinson and Gehrels (2008, p. 123–124) state, " a single detrital zircon grain cannot be used to overturn a Middle Triassic correlation for the Holbrook Member…." Perhaps the same should be said of any of the single detrital zircon ages that are coming from the Chinle Group, which contain evident inconsistencies and imprecisions.

The most recently published Chinle data from Ramezani et al. (2011) are mainly consistent with the previously reported ages and support the interpretation of a younger (i.e., ~220 Ma) Carnian-Norian boundary. These authors carefully documented the stratigraphic levels of their samples within the well-known outcrop sections at Petrified Forest National Park and obtained ages of: 225.2 Ma (for 6 samples) from near the base of the Chinle (Mesa Redondo Formation); 223 Ma (7 samples) from the middle-upper Blue Mesa Member (Petrified Forest Formation); ages for the Sonsela Sandstone range from 219.3 Ma (6 samples) in the lower part of the unit to ~218 Ma (6 samples) from the middle Sonsela. Strata near the base of the overlying Painted Desert Member of the Petrified Forest Formation yielded ages of 213.1 to 213.9 Ma (for 9 samples).

Irmis et al. (2011) also have published Chinle data recently that are relevant to this discussion. Sampling Chinle strata in northwestern New Mexico that they correlated as equivalent to the uppermost Blue Mesa Member of the Petrified Forest Formation in Arizona, Irmis et al. obtained a mean age of 218.7 Ma (from 12 grains, but removing the cluster of the three oldest dates). From the Hayden Quarry at Ghost Ranch, north-central New Mexico, they obtained a maximum possible age of 211.9 Ma. Irmis et al. (2011) assume the Hayden Quarry strata are equivalent to the lower Painted Desert Member of the Petrified Forest Formation, although they concede that no direct lithostratigraphic correlation is possible here. In both studies, the authors conclude that most of the Chinle Group strata were deposited during the Norian stage, but we reiterate that this conclusion is based almost entirely on the acceptance of the palaeomagnetic correlation of Muttoni et al. (2004), which we reject. Thus, in the broader sense, these newer dates are consistent with other recent data. Hence, our conclusion remains that the Chinle Group detrital zircon ages are consistent with a Carnian-Norian boundary of about 220 Ma, but are not by themselves a reliable dataset.

Older reports containing numerical ages either are unreliable or else consistent with a Norian base at about 220 Ma, but inconclusive (Table 1). For example, a Rb–Sr age of ~225 Ma from detrital mica in upper Chinle Group strata in West Texas (Long and Lehman, 1993) is stratigraphically high in the Revueltian, and thus clearly Norian, but this date is too old to be consistent with the detrital zircon ages. A U/Pb age of 225 ± 3 Ma on detrital zircon in rhyolites of earliest Norian age in Alaska (Gehrels et al., 1987) is an age based on old analytical techniques and therefore imprecise. A U–Pb age of 212 ± 2 Ma on calcrete calcite from the New Haven Arkose of the Newark Supergroup (Wang et al., 1998) offers only a very young constraint on the numerical age of the Carnian–Norian boundary because the calcretes lie far above the base of the Norian.

One possibly reliable numerical age is that of ~231 Ma reported by Furin et al. (2006) from the Pignola 2 section in Italy. There, an ash bed occurs about 2 m above beds interpreted as representing the "Carnian pluvial event" (Furin et al., 2006; Rigo et al., 2007) and is of likely early Tuvalian age, as dated by the LO of the conodont *Metapolygnathus nodosus* and a palynomorph assemblage including typical Carnian forms such as *P. densus*, *C. secatus*, *Ovalipollis pseudoalatus* and the characteristic Tuvalian form *P. quadruplicis*. We note, however, that the age of the Carnian pluvial event is documented by most workers as early Carnian (latest Julian) in age (Simms and Ruffell, 1990; Simms et al., 1994; Kozur and Bachmann, 2010a; Roghi et al., 2010). Therefore, like Mundil et al. (2010) we believe the Pignola 2 age of ~231 Ma sets only a maximum age for the base of the Norian, and this age also represents an estimate of the age of the boundary between the Julian and Tuvalian substages (=early-late Carnian boundary).

An Ar/Ar age of ~228 Ma from a tuff in the nonmarine Ischigualasto Formation in Argentina (Rogers et al., 1993), with a 1% correction that would translate it to a normalized U–Pb age (Jourdan et al., 2009; Mundil et al., 2010), is indistinguishable from the Pignola 2 date. The Argentinian age is just below a stratigraphic section that yields Adamanian (late Carnian) tetrapods, and this supports a correlation of the beginning of the Adamanian with the beginning of the Tuvalian (Lucas, 1998, 2010c). A recently published ^{40}Ar/^{39}Ar date from the uppermost Ischigualasto Formation of 217.0 ± 1.7 Ma (Shipman, 2004; Currie et al., 2009) is at the top of the Adamanian section and translates to ~220 Ma with a 1% correction (Table 1). Therefore, on face value, this suggests an end-Adamanian age (~ Carnian–Norian boundary) of about 220 Ma.

In conclusion, the evidence presented above shows that none of the most reliable existing numerical ages support a long Norian Stage. Instead, these ages all are much more consistent with a Carnian–Norian boundary at around 220 Ma. Precise and reliable ages are still needed, however, to unequivocally establish the precise numerical age of the base of the Norian, which may well prove to be slightly younger or slightly older than 220 Ma.

8. Conclusion: age and correlation of the Carnian–Norian boundary

Upon consideration of all available data, we offer the following conclusions:

1. Three biostratigraphic datasets (palynomorphs, conchostracans and tetrapods) all reliably identify the same position (within the normal range of biostratigraphic resolution) of the Carnian–Norian boundary in nonmarine strata of the Chinle Group (American Southwest), the Newark Supergroup (eastern USA–Canada) and the German Keuper. The stratigraphic placement of the boundary in these strata can be correlated with confidence to the marine record.
2. These biostratigraphic datasets place the Carnian–Norian boundary at the base of the Warford Member of the lower Passaic Formation in the Newark Basin (vertebrates and conchostracans), and by sporomorphs between the Lockatong/New Oxford and Lower Passaic/Heidlersburg palynological associations (Cornet, 1977; Cornet and Olsen, 1985; Litwin et al., 1991). This boundary is always shown as slighty diachronous (slightly different position in different sections) between member CD (one member below the Warford Member) and the base of the Graters Member (one member above the top of the Warford Member). It includes the conchostracan/tetrapod boundary and is also situated within the lower Passaic Member. As was widely accepted prior to 2002, this places the base of the Norian in the Newark Supergroup somewhere within magnetozone E13. This corresponds to a numerical age of ~217–218 Ma based on the Newark cyclostratigraphy if we assume that there is no Rhaetian unconformity in this section. Under the assumption of a 4 myr long

Rhaetian and only 200,000 years of Rhaetian time preserved in the Newark Supergroup, we arrive by counting 400 ka McLaughlin cycles at a Norian base of 221.5 Ma.

3. Therefore, the magnetostratigraphic correlation, e.g., by Krystyn et al. (2002), Gallet et al. (2003, 2007) and Muttoni et al. (2004, 2010), that places the Norian base much lower in the Newark Basin (~ lower Stockton, base of magnetozone E7, ~229 Ma: Gallet et al., 2007; close to the base of the Lockatong Formation, correlative to Newark magnetozone E7r: Muttoni et al., 2004; or base of E8: Muttoni et al., 2010) must be rejected. These correlations became the justification for placing the Carnian–Norian boundary at ~228 Ma or at ~229 Ma, based on the Newark cyclostratigraphy. The "long Norian" thereby created must be abandoned.

4. No other data have been forthcoming to support a Carnian–Norian boundary anywhere near as old as 228 Ma. Instead, all reliable existing radioisotopic age data are consistent with a Norian base at ~220 Ma.

5. Although the Carnian–Norian boundary can be placed confidently in the vicinity of 220 Ma, more reliable and precise radioisotopic ages are needed to assign a precise age to the Carnian–Norian boundary.

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