*Annu. Rev. Earth Planet. Sci. 1997. 25:337–401*


# STRATIGRAPHIC RECORD OF THE EARLY MESOZOIC BREAKUP OF PANGEA IN THE LAURASIA-GONDWANA RIFT SYSTEM

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KEY WORDS: Triassic, Jurassic, lakes, Milankovitch cycles, basalts

ABSTRACT

Rift basins of the Central Atlantic Margins (CAM) of North America and Morocco preserve largely continental sequences of sedimentary strata and less important minor basalt flows spanning much of the early Mesozoic. The best known is the Newark basin of New Jersey, New York, and Pennsylvania where an astronomically calibrated magnetic polarity time scale is developed.

Lacustrine cycles of Milankovitch origin are commonly present in CAM basins, with the period changing from 10 ky (paleoequator with coals), to 20 ky (4°–10°N), to perhaps 40 ky northward with evaporites. Cycles of ∼100 ky, 413 ky, and ∼2 my are also important.

Four mostly unconformity-bounded tectonostratigraphic sequences are present. The Anisian TS I is fluvial and eolian. TS II–TS IV (Late Triassic to Early Jurassic) consist of "tripartite" lacustrine sequences caused by extension pulses. The Newark basin accumulation rate history allows comparison with quantitative rift basin models.

The North American plate's slow northward drift resulted in a relative shift of climate, although the rapid humidification during the latest Triassic and Early Jurassic is associated with a sea-level rise. The Triassic-Jurassic mass extinction is of independent origin, plausibly impact related.



0084-6597/97/0515-0337$08.00

## INTRODUCTION

During the Triassic and Early Jurassic, an enormous series of rift basins formed along a wide zone extending from the future Gulf of Mexico through Nova Scotia, Morocco, the Tethyan margin, western Europe, Greenland, and Spitzbergen (Figure 1). This paper is a review of the stratigraphic records of the Central Atlantic Margin (CAM) component of these rifts (Figures 2, 3). Specifically, these rifts include the following: 1. the exposed remnants of rifts in Eastern North America containing rocks termed the Newark Supergroup (Figure 1) (Olsen 1978, Froelich & Olsen 1984; see bibliography in Margolis et al 1986) and their equivalents preserved beneath the Atlantic Coastal Plain and on the continental shelf; 2. the rift sequences present in the subsurface along the margin of the Gulf of Mexico; and 3. the exposed and subsurface basins in Morocco. Related sequences occur in Iberia, Central Europe, Great Britain, Norway, Greenland, and Svalbard. The mostly continental half graben and half graben complexes filled with thousands of meters of continental sedimentary and igneous strata provide a uniquely detailed record of early Mesozoic climatic, tectonic, and evolutionary patterns and processes over an interval of time lasting more than 30 million years (Figure 3).

A key point is that the dramatic cyclicity of the rift-basin lacustrine sequences allows for the quantitative analysis of the varying and intertwined effects of tectonics, plate position, and climate. The large latitudinal spread over which these basins developed, covering about 30$^\circ$ of paleolatitude, allows an unusually detailed look at the Late Triassic and earliest Jurassic world over a broad transect of climate and tectonic processes.

There are six major themes that this review of CAM rift basin stratigraphic evolution will cover: 1. sedimentary and climatic cyclicity; 2. tectonic control of basin facies and sequences; 3. identification of tectonostratigraphic sequences; 4. linkages between presently isolated depositional basins; 5. effects of the drift of the North American plate into different climate belts; and 6. regional and global climate events.

## SEDIMENTARY AND CLIMATIC CYCLICITY

Perhaps the most striking aspect of CAM basin sequences is the prevalence of cyclicity in lacustrine lithologies evident at many thickness scales (Figures 4, 5). This sedimentary cyclicity was primarily controlled by lake-level variations following quasi-periodic Milankovitch-type climate cycles governed by celestial mechanics (Hays et al 1976). The careful and detailed review by Smoot (1991a) explains the climatic significance of various sedimentary facies and cycles within the CAM basins. Although cyclicity is pervasive in CAM basins,

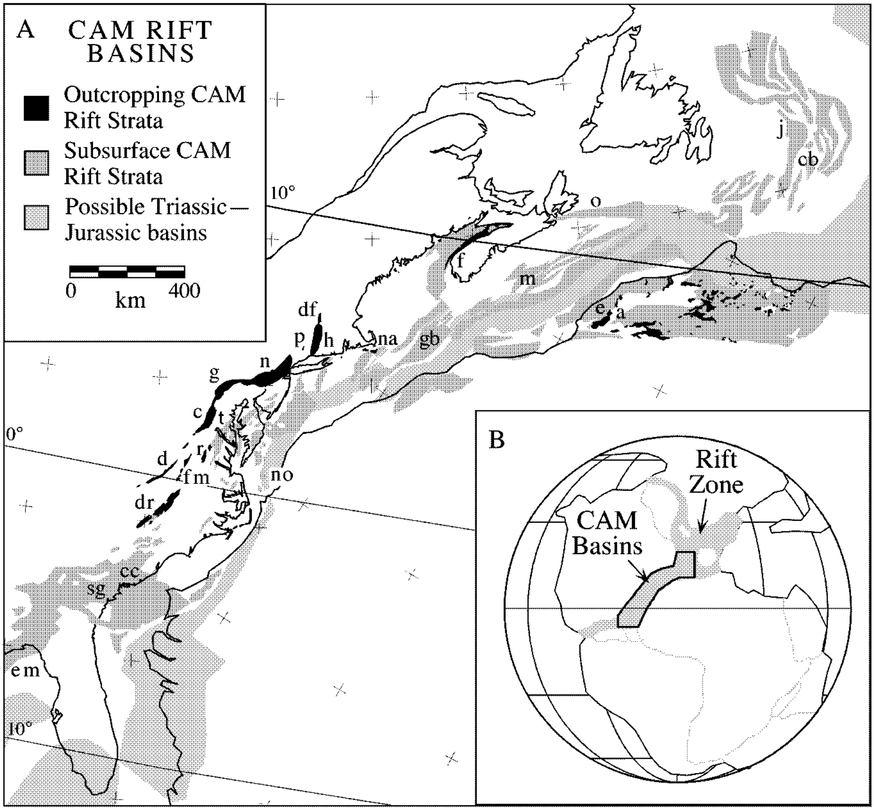


*Figure 1* Central Atlantic Margin (CAM) rift basins in their paleogeographic position for the Carnian (~225 Ma), based on Newark basin data in Kent et al (1995). Positions of paleolatitude lines have an estimated error of ±2°. (*A*) CAM basins: a, Argana basin; c, Culpeper basin; cb, Carson basin; cc, Clubhouse Crossroads area in South Georgia basin; d, Dan River–Danville basins; df, Deerfield basin; dr, Deep River basin; e, Essaouira basin; em, eastern most basins with Eagle Mills Formation; f, Fundy basin; fm, Farmville and related basins; g, Gettysburg basin; gb, Georges Bank basin; h, Hartford basin; j, Jeanne d'Arc basin; m, Mohican basin; n, Newark basin; na, Nantucket basin; no, Norfolk basin; o, Orpheus basin; p, Pomperaug basin; t, Taylorsville basin. Sources for map are Olsen et al (1989), Holser et al (1988), Tankard & Balkwill (1989), Heyman (1989), Hutchinson & Klitgord (1988a,b), Benson (1992), and Beauchamp (1988). Strata in outcropping basins labeled o, f, df, h, p, n, g, c, t, r, fm, d, and dr are the major components of the Newark Supergroup. (*B*) Reconstruction of Pangea for the Carnian, showing the rifting zone (*gray*) and the CAM basins detailed in (*A*).

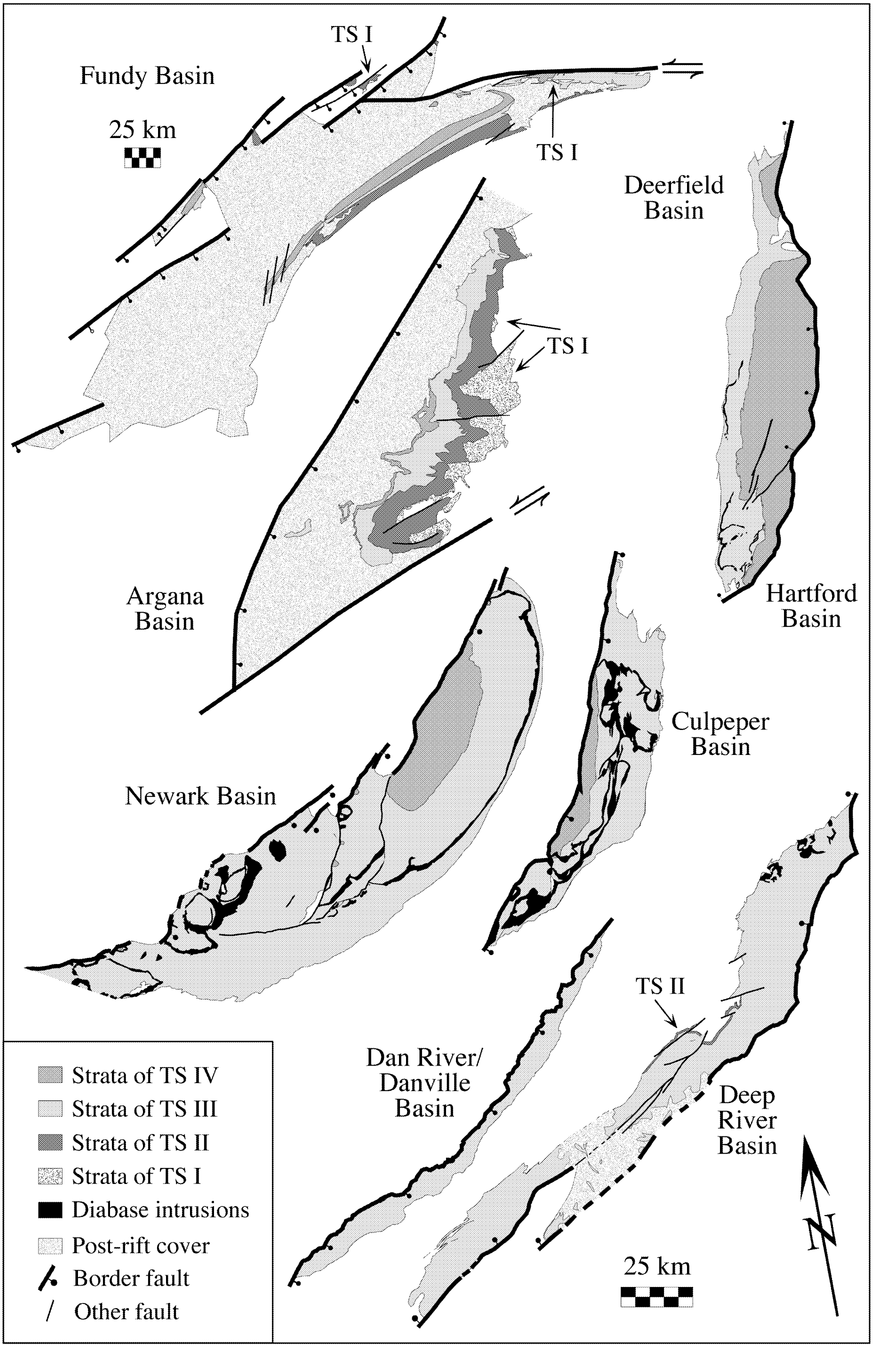


*Figure 2* Maps of selected CAM basins (base maps from Anders & Schlische 1994). All the basins are drawn to the same scale except the Fundy basin.

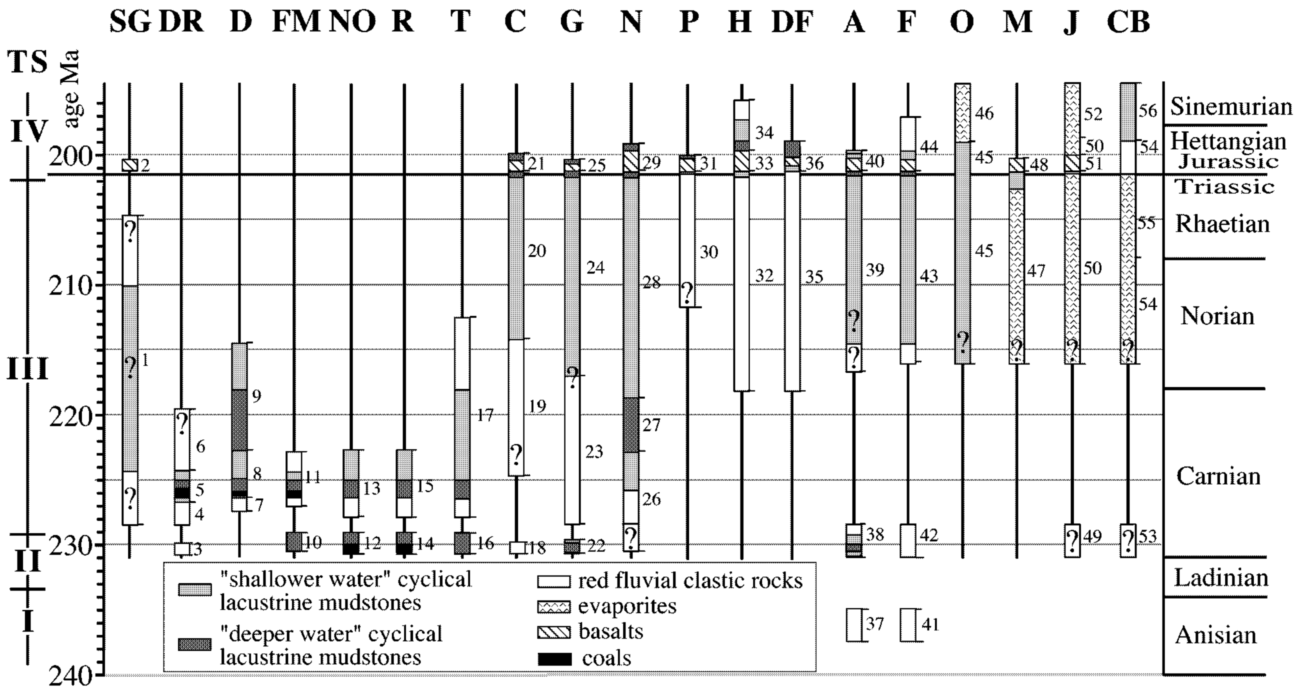


*Figure 3* Ages, basic facies, and formations in the CAM basins discussed in the text. Formation names are indicated by numbers explained in Table 1. Basins are as follows: SG, South Georgia; DR, Deep River; D, Danville/Dan River; FM, Farmville and Briery Creek; NO, Norfolk; R, Richmond; T, Taylorsville; C, Culpeper; G, Gettysburg; N, Newark; P, Pomperaug; H, Hartford; DF, Deerfield; A, Argana; F, Fundy; O, Orpheus (Eurydice P-36 well); M, Mohican (Glooscap C-63 well; Pe-Piper et al 1992); J, Jeanne d'Arc (Spoonbill C-30 well); CB, Carson (Osprey H-84 well) (O, J, and CB adapted from Holser et al 1988).

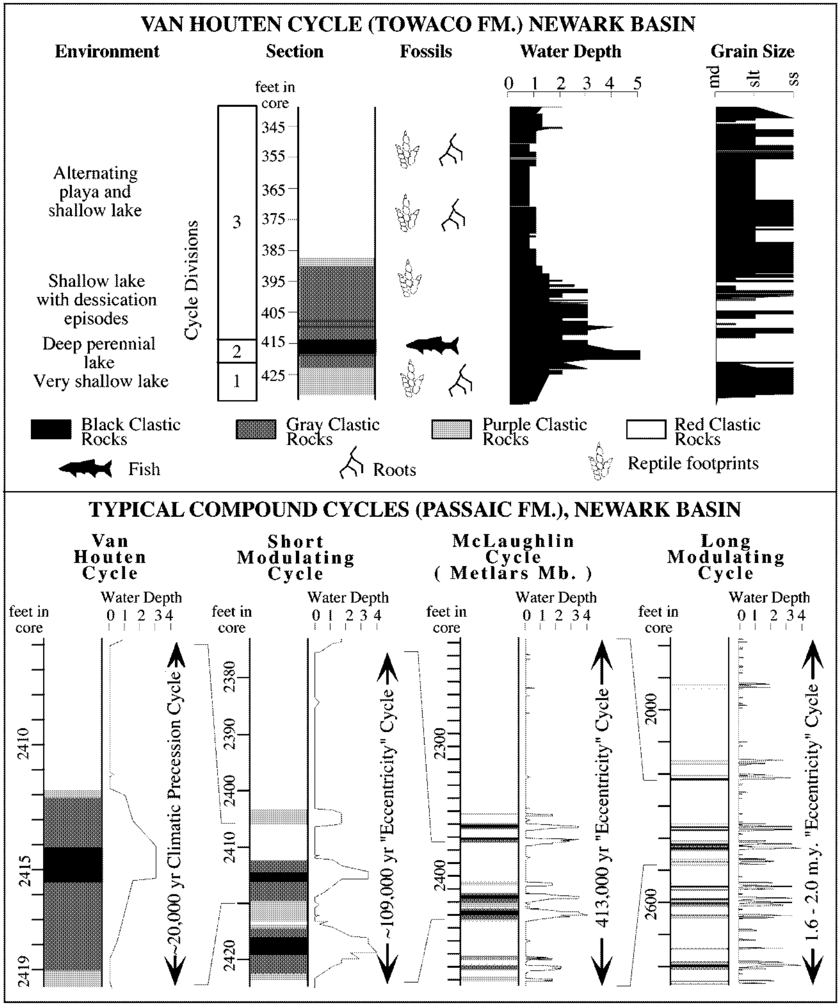


*Figure 4* Typical Van Houten cycles and compound cycles of the Newark basin (modified from Olsen et al 1996b).

the best documented example of cyclicity and periodicity is from the Newark basin of the Newark Supergroup (Figures 1, 6). Sedimentary cyclicity in the lacustrine portions of the basin was first described by McLaughlin (1933), who demonstrated the great lateral extent of the large-scale (~100 m) alternations of gray to black shales and red mudstones (McLaughlin 1933, 1946, 1959). Van Houten (1962, 1964, 1969, 1980) extensively studied the Lockatong Formation and provided a detailed look at Newark basin cyclicity, ascribing it to

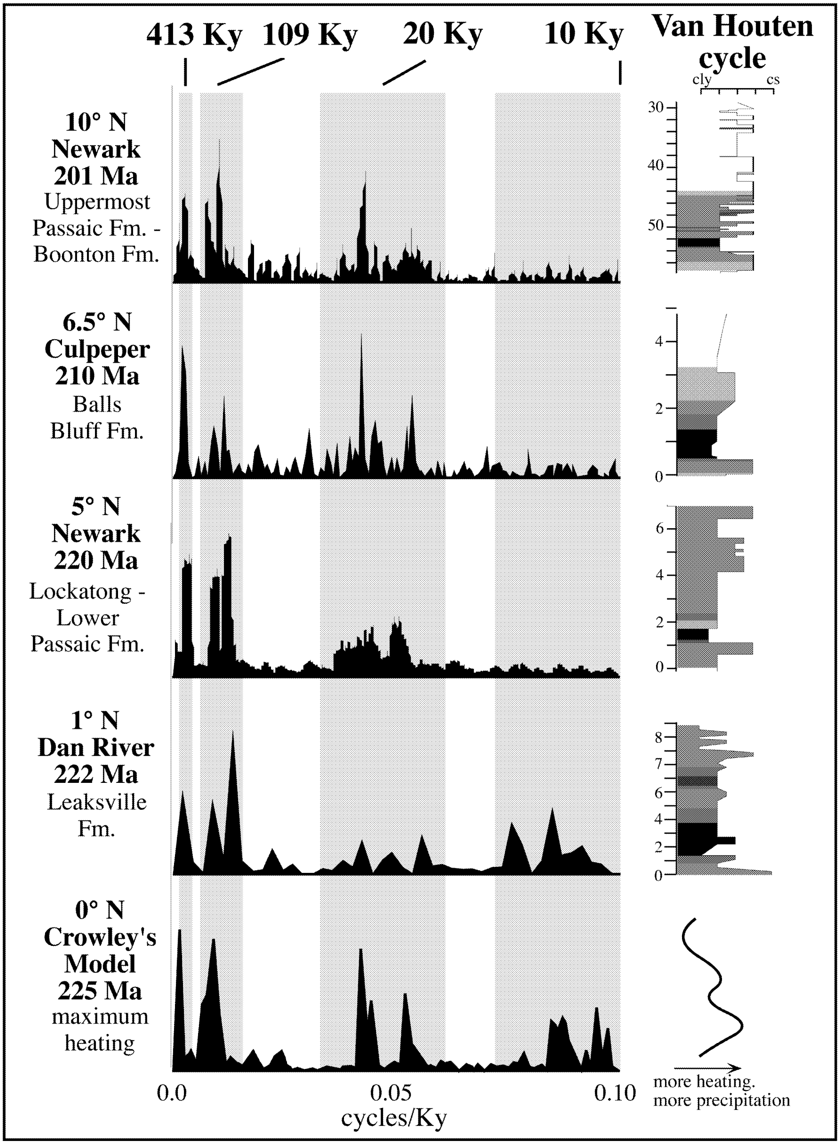


*Figure 5* Representative cycles, power spectra, interpreted periods, and theoretical insolation curve for various portions of Newark Supergroup basins. The "Crowley's Model" power spectrum is from Crowley et al (1992). [Key to lithology colors as in Figure 4 (*top panel*).]

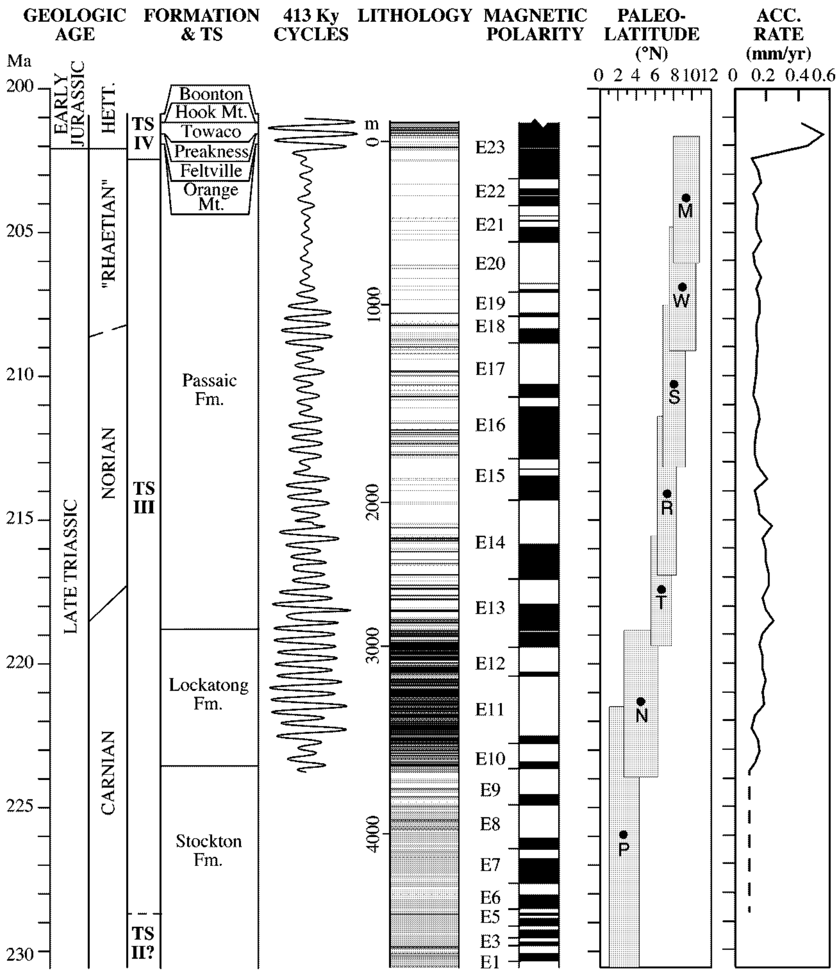


*Figure 6* The Newark basin stratigraphic record based on NBCP and ACE data (largely from Kent et al 1995, Olsen et al 1996a,b). [Key to lithology colors as in Figure 4 (*top panel*).]

climatic variations controlled by celestial mechanics. Olsen (1986), following the paradigm of Hays et al (1976) for Quaternary cyclicity, used Fourier analysis to obtain quantitative assessments of the periodicity of the cyclically bedded outcrops, thus corroborating Van Houten's hypothesis (Figure 5). Pervasive cyclicity in most of the Newark basin section was confirmed by continuous coring of the Newark basin by the Newark Basin Coring Project (NBCP) (Olsen et al 1996a, Olsen & Kent 1996) and the Army Corps of Engineers (ACE)

(Fedosh & Smoot 1988, Olsen et al 1996b). Together, over 12 km of core from these projects cover virtually all of the approximately 7 km–thick Newark basin section, including more than 5 km of cyclical lacustrine rocks that show patterns consistent with the Milankovitch theory of climate change (see Smoot & Olsen 1994, Olsen & Kent 1996) (Figure 6).

As seen in the cores and outcrops of the Newark basin lacustrine units, as well as in several other CAM basins, the fundamental sedimentary cycle is a transgressive-regressive lacustrine sequence termed a Van Houten cycle (Olsen 1986, 1990) (Figure 4). Vertical sequences of these cycles trace out three orders of modulating cycles; these cycles are expressed as variations in the degree of development of sedimentary structures indicative of subaerial exposure or deposition in deep water within successive Van Houten cycles. The short modulating cycle contains 4–6 Van Houten cycles; the McLaughlin cycle exhibits 4 short modulating cycles; and the long modulating cycle is made up of about 5 McLaughlin cycles (Figure 4). Frequency analysis of sections exhibiting this cyclicity by Fourier and moving window ("evolutive") techniques indicates that the cyclicity conforms to the expectations of Milankovitch climate theory for the tropics. Specifically, the climate changes that controlled lake level were governed by the cycle of climatic precession, modulated by the ∼100-ky, 413-ky, and ∼2-my so-called eccentricity cycles of the Earth's orbit.

As expected from Milankovitch theory, the expression of the insolation cycles in continental climate, and hence the rock record, varies dramatically with latitude in the CAM basins. Within specific time intervals, the distinctive frequency characteristics of the cyclicity as well as the overall facies (i.e. humid vs arid) change in a methodical way that is well explained in the Milankovitch paradigm. In theory, the region around the equator should be dominated by a "double" climatic precession cycle of ∼10 ky that gives way farther from the equator to a ∼20-ky cycle of the type seen in the Newark basin and commonly associated with the climatic precession cycle (Crowley et al 1992) (Figure 5). In addition, the middle and high latitudes should be dominated by the effects of the 41-ky cycle of the obliquity of the Earth's axis. In fact, this domination is manifest in the CAM basins (see below). This pattern of latitudinally dependent cyclicity is apparently in very good agreement with the most recent well-corroborated paleogeographic reconstruction (Witte et al 1991, Kent et al 1995, Kent & Muttoni 1996, Kent & Olsen 1996), which places Virginia at the equator in the early Late Triassic (Figures 1, 6). The change in cyclical mode is paralleled by a dramatic change in the dominant sediment type, with coals and black shales in the south giving place to aeolianites in the northern Newark Supergroup basins (Olsen et al 1989). The cyclicities so well displayed in the CAM basins provide not only a metronome for calibrating other basin phenomena, such as the magnetic polarity sequence from the basins, accumulation rates, and hence indirectly subsidence (Figure 6), but also are the first clear demonstration of the local effects of Milankovitch cycles on continental climate over a broad latitudinal swath (Figure 5).

## TECTONIC CONTROL OF BASIN FACIES AND SEQUENCES

Although cyclicity is a very prominent part of CAM basin sequences, there are larger-scale patterns such as lateral and vertical transitions in facies types that were evidently produced by tectonic processes.

As presently understood, CAM basins are largely half graben and half graben complexes. The basins are bounded on one side by a major mostly normal fault system, towards which the basin strata tilt (Figure 7). There is considerable evidence that the fault systems that border the present half graben of the CAM basins were active during sedimentation (see Schlische 1993). Most important are the locally derived conglomerates at the fault-bounded sides of the basin. Although these conglomerates do not comprise a volumetrically large part of the basin fill, they do indicate that a local source of relief was maintained throughout the basin's sedimentological history along the faulted margin (Longwell 1937, Schlische 1993).

It was once thought that episodes of faulting or protracted periods of accelerated faulting would produce a progradation of alluvial fans from the border fault because of the increase in basin relief (Barrell 1917, Lorenz 1988, Ressetar &

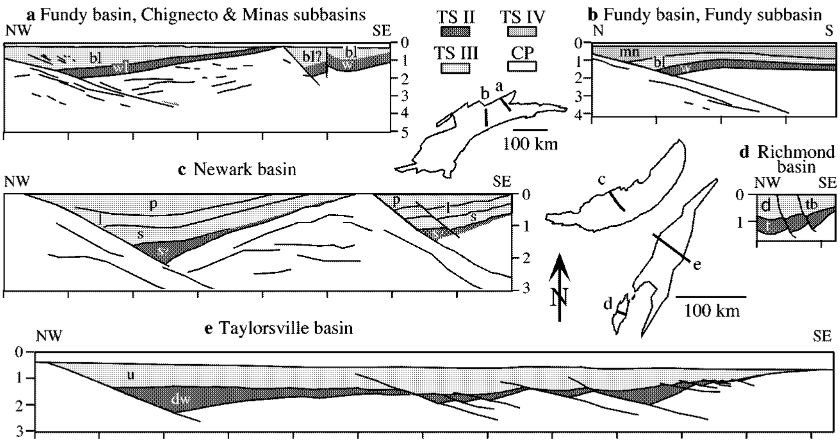


*Figure 7* Representative cross sections of basins based on industry seismic profiles. Profile lines *a* (Fundy basin and Chignecto and Minas subbasins) and *b* (Fundy basin and subbasin) are adapted from Withjack et al (1995); *c* (Newark basin), revised interpretation based on Costain & Coruh (1989); *d* (Richmond basin), adapted from Cornet & Olsen (1990); and *e* (Taylorsville basin), based on a variety of unpublished seismic profiles and Milici et al (1991).

Taylor 1988, Mickus et al 1988). However, recently proposed models of the sedimentological effects of faulting within half graben argue that an increase in faulting would produce a retreat of alluvial fans owing to lakes ponding against the tilting depositional surface and the increase in accommodation space (Blair & Bilodeau 1988). In fact, the relationship between alluvial fan and lacustrine deposits in the Newark basin suggests that in most cases fans prograde during overall dry times irrespective of faulting (Olsen et al 1989).

There is also ample evidence that the basins, as presently delineated, were, at least in part, half graben during deposition (Schlische 1993, Withjack et al 1995). Synsedimentary half graben development is evidenced by the following: 1. decreasing dip of strata up-section both in map view and in drill holes; 2. fanning of strata toward the border fault system as seen in seismic profiles (Hutchinson & Klitgord 1988a, Costain & Coruh 1989) (Figure 7); 3. thickening of intervals of strata, particularly sedimentary cycles, toward the border fault as derived from continuous core and outcrop (Olsen et al 1996a); and 4. the results of provenance and paleocurrent studies (Klein 1963, Glaeser 1966, Weddle & Hubert 1983). Because of the half graben geometry of the basins, sedimentary sequences generally coarsen downward at the edges of basins, as seen in drill holes, and coarsen towards the basin margins in map view, as pointed out by McLaughlin & Willard (1949).

All of the CAM basins have sequences that show strong vertical changes in overall facies reflecting an interplay of large-scale tectonic and climatic changes with time. The basic outline of these has long been clear (Van Houten 1977, Manspeizer 1988). In specific, the basin sections are made up of one or more tripartite sequences consisting of a lower, relatively coarse-grained, fluvial facies with a rapid transition upwards into relatively fine-grained, deeper-water facies, followed by a slower transition upward into shallower-water and even coarse fluvial facies (Olsen et al 1989) (Figures 3, 6). Similar asymmetrical vertical sequences have been observed in many other continental rifts (Lambiase 1990). Such sequences have been explained as the result of a specific history of tectonic events (Manspeizer & deBoer 1989), as a consequence of sequential filling of hydrographically linked basins (Lambiase 1990), and as the necessary result of the filling of a widening half graben bounded by a normal fault (Schlische & Olsen 1990, Schlische 1993, Contreras et al 1996).

Schlische & Olsen (1990) proposed a simple quantitative model in which the rate of subsidence, the rate of inflow of sediment into the basin, and the basic geometry of the basement components of the basin remain constant (Olsen 1990). As a simple consequence of the increase of the area of the depositional surface though time, the full tripartite lacustrine sequence is produced. This basin-filling model predicts the sediment accumulation rate history of the basin; once lacustrine conditions are established, accumulation rate should decline exponentially.

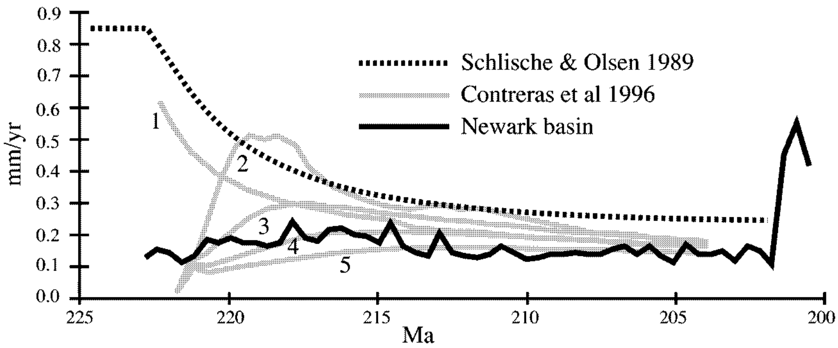


*Figure 8* Comparison of accumulation rate predictions of models of rift basin evolution with data from the Newark basin (see Figure 6) of various authors. Numbers 1–5 refer to predictions at increasing distances from the border fault of Contrerar et al (1996).

The first quantitative data on long-term sedimentation rates comes from the NBCP and ACE cores (Olsen et al 1996a,b) (Figure 6). Comparison of the sedimentation rates of the Triassic-age part of the basin section with the predictions of the Schlische & Olsen (1990) model shows some first-order differences (Figure 8). Specifically, a simple exponential decline in sedimentation rates is not seen; however, the facies transitions predicted in the Schlische & Olsen model are confirmed.

A much more realistic, although more complicated, model of evolving rift-basin geometry is presented by Contreras et al (1996) (Figure 8). This is a three-dimensional finite-difference model based on the fault-growth models of Cowie & Scholz (1992). Contreras et al (1996) show that for a half graben bounded by a single normal fault the accumulation-rate curve predicted by the Schlische & Olsen (1990) model occurs only at the depocenter of the basin, adjacent to the border fault. All other parts of the basin have different accumulation-rate curves. An accumulation-rate curve similar to that seen in the Newark basin data (Figure 6) is obtained in the model if the basin is sampled by laterally offset drill-hole data as were obtained in the coring project.

Viewed in light of the above models, each tripartite sequence probably represents a different tectonic episode, basically a rejuvenation of extension or an acceleration of extension. This hypothesis can be expressed as an extension of the Schlische & Olsen (1990) or Contreras et al (1996) models. Thus, relatively low rates of extension and basin subsidence allow the available sediments to fill the basin to the outlet, resulting in a hydraulically open basin with through-flowing streams and a coarse-grained basin fill. Slow rates of extension should result in similarly slow rates of footwall uplift. Hence, it could have been relatively easy for inflowing or outflowing rivers to breach the footwall. The fluvial portions of the Stockton Formation of the Newark basin (Smoot 1991a,b) may be an example of streams leaving through the footwall, whereas the New Haven Formation of the Hartford basin (Hubert et al 1978) may be an example of a fluvial system entering the basin though the footwall (Figure 2). When extension rates increased, basin floor subsidence and tilting increased. As a consequence, the basin expanded at a rate faster than it could fill, and a closed drainage basin resulted. When sufficient water became available, a lake could form. Smoot (1991a), on the other hand, feels that an autogenic reduction of fluvial gradient due to clogging of the outlet resulted in the fluvial-lacustrine transitions. As long as extension rates remained high, the distance between the outlet and the basin floor would continue to increase as the area of the depositional surface expanded; and as long as there was sufficient precipitation, lacustrine conditions would be maintained (Schlische & Olsen 1990). High extension rates would lead to high rates of footwall uplift, and the basin would be sourced from either the hanging wall [upper Stockton and Lockatong formations, Newark basin (Glaeser 1966)] or the axis [Passaic Formation, Newark basin (Smoot 1991a)].

As extension decreased, the basin would tend to fill, eventually reestablishing fluvial conditions. Footwall uplift would correspondingly decrease and the footwall could again be breached by streams, perhaps capturing large drainage areas on the back of the footwall uplifts (upper Portland Formation, Hartford basin; McInerney 1993) (Figure 2). If subsidence stayed low or stopped, the outlet would be cut into, and eventually the basin fill would begin to erode. Thus, the full cycle of an extension pulse would tend to produce one of the tripartite lacustrine sequences. If the magnitude of the extension pulse is relatively small, then a short-lived excursion though a tripartite sequence might be produced. If the pulse of accelerated extension followed a period of slow extension or erosion or if it was of large enough magnitude to produce erosion of footwall sediments before they could be onlapped by the new tripartite sequence, then an unconformity might develop that would separate the two sequences along the up-dip portions of the basin. However, as long as the basin remained hydraulically closed, a conformable relationship between two successive sequences would still be expected at the depocenter of the basin. In addition, an abbreviated yet strong pulse might only produce a fluvial sequence. Multiple tripartite sequences and unconformity-bound fluvial sequences are observed in many CAM basins, which suggests that their basin fill can be organized into discrete tectonostratigraphic sequences.

## TECTONOSTRATIGRAPHIC SEQUENCES

Traditionally, the main organizing principle within basin sequences has been the recognition of mappable lithostratigraphic formations defined on one or a few criteria, such as color or grain size (Figure 2; Table 1). A detailed review

**Table 1** Formations of CAM basins

| No.[a] | Formation name | Thickness | Description | References |
|---|---|---|---|---|
| 1 | Unnamed | 120 | Red clastic rocks | Gohn et al 1983 |
| 2 | Clubhouse Crossroads Basalt | 260 | Basalt flows | Gottfried et al 1983 |
| 3 | "Millstone Grit" of Pekin Fm (in part) | 100 | Gray clastic rocks with occasional coals | Reinemund 1955, Stagg et al 1985 |
| 4 | Pekin Fm | 400 | Red and gray sandstone and mudstone | Reinemund 1955, Textoris & Gore 1994 |
| 5 | Cumnock Fm | 260 | Gray and black mudstone, sandstone, and coal | Reinemund 1955, Textoris & Gore 1994 |
| 6 | Sanford Fm | 850 | Mostly red clastic rocks | Reinemund 1955, Textoris & Gore 1994 |
| 7 | Pine Hall Fm (in part) | 900 | Gray and brown conglomerate and sandstone | Thayer 1970 |
| 8 | Leaksville (in part), Cow Branch (lower mb), Pine Hall (in part), Stoneville fms (in part), and Dry Fork (in part) | 1200 | Cyclical gray and black mudstone, sandstone, and coal overlain by cyclical red and gray sandstone and mudstone | Meyertons 1963, Thayer 1970, Olsen et al 1978, Kent & Olsen 1996 |
| 9 | Leaksville (in part), Cow Branch (upper mb), and Stoneville fms (in part) | 3000 | Cyclical gray and black mudstone and sandstone, overlain by cyclical red and gray mudstone and sandstone | Meyertons 1963, Thayer 1970, Thayer et al 1978, Kent & Olsen 1996 |
| 10 | Unnamed | 700 | Cyclical gray and black mudstone | Olsen et al 1982, Goodwin et al 1986 |
| 11 | Unnamed | 1000 | Red clastic rocks overlain by cyclical gray and black mudstone with minor coals, grading up into red and gray coarse clastic rocks | Roberts 1928, Olsen et al 1982, Goodwin et al 1986 |
| 12 | Unnamed | 3000 | Seismic character suggests clastic with interbedded coal | Musser 1993 |

| | | | | |
|---|---|---|---|---|
| 13 | Unnamed | 2500 | Seismic character suggests gray and black sequence bracketed above and below by red clastic rocks | Musser 1993 |
| 14 | Tuckahoe Fm | 1000 | Gray sandstone and conglomerate, overlain by coals and gray sandstone, overlain by cyclical black and gray mudstone | Shaler & Woodworth 1899, Olsen et al 1989, Ediger 1986, Goodwin et al 1996, Cornet & Olsen 1990 |
| 15 | Turkey Branch Fm, Otterdale Fm (in part) | 1300 | Red and gray clastic rocks grading up into cyclical mostly gray and black mudstone and sandstone | Shaler & Woodworth 1899, Cornet & Olsen 1990 |
| 16 | Doswell Fm | 2300 | Minor red and gray clastic rocks overlain by cyclical gray and black mudstone and sandstone | Weems 1980, Cornet & Olsen 1990, LeTourneau 1996 |
| 17 | Unnamed | 2000 | Red and gray clastic rocks overlain by cyclical gray and black mudstone overlain a red then a gray clastic sequences, overlain by red clastics | Cornet & Olsen 1990, LeTourneau 1996 |
| 18 | Reston Mb (Fm) | 100 | Red and gray conglomerate | Lee & Froelich 1989, Smoot 1991a |
| 19 | Manassas Fm | 1000 | Red and gray sandstone and minor mudstone | Roberts 1928, Lee & Froelich 1989, Smoot 1991a |
| 20 | Balls Bluff, Catharpin Creek,[b] Tibbstown fms (lateral equivalent) | 2000 | Cyclical red, gray, and black mudstone and sandstone, grades laterally into red and gray conglomerate | Lee & Froelich 1989, Olsen et al 1989, Gore 1988c, Smoot 1991a |
| 21 | Mt. Zion Church Basalt, Midland Fm, Hickory Grove Basalt, Turkey Run Fm, Sander Basalt, Waterfall Fm[b] | 3000 | Cyclical red, gray, and black mudstone and minor carbonate (Midland Fm) with three interbedded basalt flow units | Lindholm 1979, Hentz 1985, Olsen et al 1982, 1989, Gore 1988c, Roberts 1989 |

*(Continued)*

**Table 1** (*Continued*)

| No.[a] | Formation name | Thickness | Description | References |
|---|---|---|---|---|
| 22 | Unnamed | 100 | Black and gray mudstone and conglomerate | Cloos & Pettijohn 1973, Cornet 1977, Smoot 1991a |
| 23 | New Oxford Fm | 2100 | Gray and red clastic rocks | Stose & Jonas 1939, Cornet 1977, Olsen et al 1982, Smoot 1991a |
| 24 | Gettysburg Fm, Hammer Creek Fm | 4900 | Cyclical red, gray, and black mudstone and sandstone, grades laterally into red conglomerate (Hammer Creek Fm) | Stose & Jonas 1939, Cornet 1977, Smoot 1991a |
| 25 | Aspers Basalt and unnamed | 60 | Basalt flows and overlying red and gray mudstone | Stose & Bascom 1929, Cornet 1977 |
| 26 | Stockton Fm[c] | 1000 | Gray, red, and minor black clastic rocks | Kümmel 1997, Smoot 1991a, Olsen et al 1996 |
| 27 | Lockatong Fm[c] | 760 | Cyclical gray, black, and minor red mudstone | Kümmel 1997, Van Houten 1960, Smoot 1991a, Olsen et al 1996 |
| 28 | Passaic Fm,[c] Hammer Creek Fm | 2700 | Cyclical red, gray, and black mudstone grading laterally into coarse red clastics (Hammer Creek Fm) | Kümmel 1997, Van Houten 1960, Smoot 1991a, Olsen et al 1996a |
| 29 | Orange Mt. Basalt, Feltville Fm, Preakness Basalt, Towaco Fm, Hook Mt. Basalt, and Boonton Fm[b] | 2000 | Cyclical red, gray, and black mudstone, sandstone, and conglomerate with minor carbonate in Feltville Fm, interbedded with three lava flow formations | Olsen 1981a,b, Smoot 1991a, Olsen et al 1996a,b |
| 30 | South Britain Fm | 250 | Red mudstone, sandstone, and conglomerate | Hobbs 1901, Weddle & Hubert 1983 |

| | | | | |
|---|---|---|---|---|
| 31 | Unnamed | 150 | Cyclical red, gray, and black mudstone, sandstone, and conglomerate, with minor carbonate in lowest sedimentary unit, interbedded with at least two lava flow formations | Olsen et al 1982 |
| 32 | New Haven Fm. | 2300 | Red and buff conglomerate, sandstone, and mudstone | Krynine 1950, Hubert et al 1978, 1992, Smoot 1991a, McInerney 1993 |
| 33 | Talcott Basalt, Shuttle Meadow Fm, Holyoke Basalt, East Berlin Fm, and Hampden Basalt[b] | 550 | Cyclical red, gray, and black mudstone, sandstone, and conglomerate, with minor carbonate in Feltville Fm, interbedded with three lava flow formations | Krynine 1950, Hubert et al 1978, 1992, Smoot 1991a, Gierlowski-Kordesch & Rust 1994 |
| 34 | Portland Fm | 2000 | Cyclical red, gray, and black mudstone, sandstone, and conglomerate, grading upward into coarse red clastics | Krynine 1950, Olsen et al 1989, Smoot 1991a, McInerney 1993 |
| 35 | Sugarloaf Fm | 1700 | Red and buff conglomerate, sandstone, and minor mudstone | Emerson 1891, Stevens & Hubert 1980, Robinson & Luttrell 1985 |
| 36 | Fall River beds, Deerfield Basalt, Turners Falls Fm,[b] and Mt. Toby Fm | 2100 | Cyclical red, gray, and black mudstone, sandstone, and minor conglomerate, grading laterally into coarse clastics (Mt. Toby Fm) | |
| 37 | Ikakern Fm | 3500 | Red and brown conglomerate, sandstone, and minor mudstone | Duffaud et al 1966, Tixeront 1977 |
| 38 | Timezgadiwine Fm | 1500 | Red conglomerate (t3) grading up into cyclical gray, purple, and red mudstone, minor carbonate, and sandstone (t4), grading up into cyclical red and brown mudstone and sandstone (t5) | Duffaud et al 1966, Tixeront 1977 |

(*Continued*)

**Table 1** (*Continued*)

| No.[a] | Formation name | Thickness | Description | References |
|---|---|---|---|---|
| 39 | Bigoudine Fm | 1300 | Red and brown conglomerate overlain by brown sandstone with minor red mudstone and halite interbeds (t6), overlain by cyclical red, gray, and black mudstone and sandstone with nodular gypsum (t7) grading up into red cyclical mudstone with nodular gypsum (t8) | Duffaud et al 1966, Tixeront 1977, Smoot 1991a |
| 40 | Argana Basalt & unnamed[b] | 300 | Basalt flows and overlying cyclical red gray and black mudstone carbonate, and sandstone | Tixeront 1977, Fiechtner et al 1992, Smoot 1991a |
| 41 | Honeycomb Pt. Fm, Economy beds of "Wolfville" Fm | 1000 | Red and brown sandstone, mudstone, and conglomerate | Nadon & Middleton 1985, Olsen et al 1989 |
| 42 | "Wolfville," Quaco, Echo Cove, and Lepreau (?) fms | 1600 | Red and gray conglomerate, sandstone, and minor mudstone | Powers 1916, Wright & Clements 1943, Klein 1962, Nadon & Middleton 1985, Hubert & Florenza 1988 |
| 43 | Red Head beds of "Wolfville" Fm, Blomidon Fm[b] | 400 | Brown sandstone with minor mudstone and conglomerate overlain by cyclical red mudstone | Powers 1916, Klein 1962, Hubert & Mertz 1984, Olsen et al 1989, Smoot 1991a |
| 44 | North Mt. Basalt, Scots Bay plus McCoy Brook fms | 500 | Basalt flows overlain by white, green, and red limestone and chert (Scots Bay) and red and brown sandstone and mudstone (McCoy Brook Fm) | Powers 1916, Klein 1962, De Wet & Hubert 1989 |

| | | | | |
|---|---|---|---|---|
| 45 | Eurydice and Chedabucto fms | 560 | Red mudstone, sandstone, and conglomerate | Lyngberg 1984, Klein 1962 |
| 46 | Argo Fm | 1600 | Halite with minor anhydrite, red sandstone, and dolomite | Holser et al 1988 |
| 47 | Argo Fm | 150 | Halite with minor anhydrite, red sandstone, and dolomite (50-m red shale unit at top) | Pe-Piper et al 1992 |
| 48 | Unnamed | 150 | Basalt flows | Pe-Piper et al 1992 |
| 49 | Kettle red beds | 250 | Red mudstone and sandstone | Holser et al 1988 |
| 50 | Osprey Fm | 1000 | Halite with minor beds of red mudstone and traces of anhydrite (90-m red shale sequence capped by dolomite just below basalt) | Holser et al 1988, Pe-Piper et al 1992 |
| 51 | Unnamed | 50 | Basalt flows | Holser et al 1988, Pe-Piper et al 1992 |
| 52 | Iroquois Fm | 300 | Gray carbonate and mudstone with minor anhydrite and halite | Holser et al 1988 |
| 53 | Kettle red beds | 180 | Red mudstone and sandstone | Holser et al 1988 |
| 54 | Osprey Fm | 1600 | Halite with minor beds of red mudstone and traces dolomite | Holser et al 1988 |
| 55 | Argo Fm | 400 | Halite with minor anhydrite, red sandstone | Holser et al 1988 |
| 56 | Iroquois Fm | 300 | Gray limestone and dolomite with minor mudstone | Holser et al 1988 |

[a]Numbers indicate units shown in Figure 3.
[b]Preceding units listed in order of superposition.
[c]Thickness based on NBCP core composite (Olsen et al 1996).

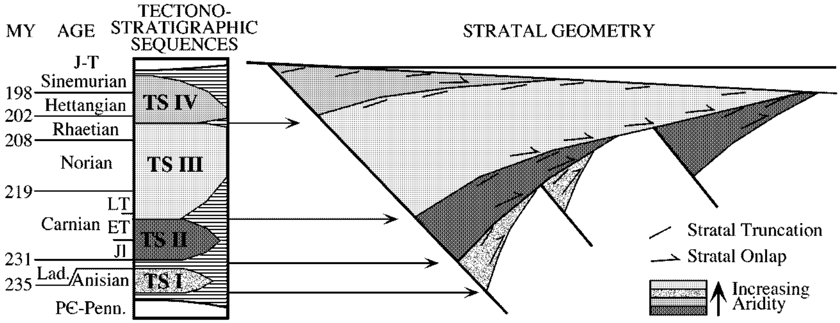


*Figure 9* Basic tectonostratigraphic sequences of CAM basins based largely on seismic profiles and coring.

of this nomenclature for the Newark Supergroup is provided by Lutrell (1989). However, recent industry and scientific drilling and seismic profiling, as well as reexamination of outcrops, has shown that four basic tectonostratigraphic sequences (TS) can be recognized that extend over most, if not all, of the CAM basins (Figure 9). These are sequences in the sequence stratigraphic sense (see Christie-Blick & Driscoll 1995); they are mostly but not always unconformity bound and probably resulted from significant changes in the rate of extension (rather than sea level changes, which in these basins is nearly irrelevant). Age is the criterion used here to group sequences in isolated basins into individual categories. These categories are as follows: TS I, initial synrift sedimentary rocks of Anisian (early Middle Triassic) age; TS II, early synrift sedimentary rocks of possibly Ladinian and Early to Late Carnian age (late Middle and early Late Triassic) age; TS III, middle synrift sedimentary rocks of Late Carnian, Norian, Rhaetian, and earliest Hettangian (Late Triassic to earliest Jurassic) age; and TS IV, late synrift sedimentary and volcanic rocks of Hettangian (Early Jurassic) age, plus the overlying late synrift sedimentary rocks of Hettangian to possibly Pliensbachian (Early Jurassic) age (Figures 3, 9). In many cases these tectonostratigraphic sequences conform to existing formal lithostratigraphic nomenclature frameworks (see below). In other cases they do not. These tectonostratigraphic sequences provide the framework for understanding the large-scale tectonic events that influenced basin evolution as well as a logical structure for the discussion of the rift sequences themselves.

TECTONOSTRATIGRAPHIC SEQUENCE I The TS I sequence comprises initial synrift sedimentary rocks that consist of strongly wedge-shaped sequences in small basins with beds fanning markedly towards faults (Figure 9). The TS I has

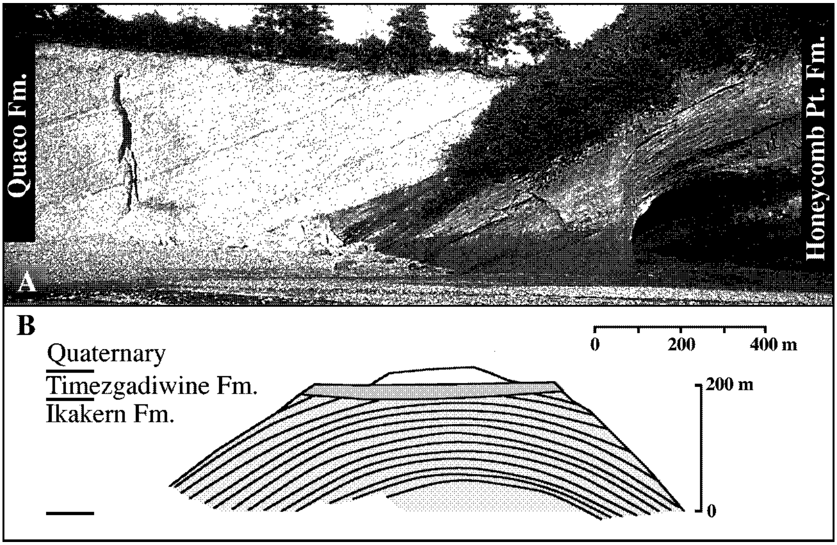


*Figure 10* Unconformities between TS I and TS II in the Fundy (A) and Argana (B) basins. (*A*) Outcrop of TS I (fluvial Honeycomb Pt. Formation, *right*) and TS II (fluvial Quaco Formation, *left*) at St. Martins, New Brunswick, Canada. Note the gentle angular unconformity with beds on the left dipping slightly less steeply than those on right. (*B*) Sketch of outcrop of TS I (Ikakern Formation) and TS II (Timezgadiwine Formation) redrawn from Tixeront (1973).

been recognized so far only in the Fundy basin and the Argana basin of Morocco. The sequence rests with a profound unconformity on underlying prerift rocks and is separated from younger synrift strata by an angular unconformity. In the limited areas where it has been recognized, TS I consists mostly of fluvial and aeolian rocks deposited under an overall arid environment. The best-known and type area is in the Fundy basin of Maritime Canada. A useful review of the early Mesozoic rocks of the Maritime Provinces of Canada is given by Greenough (1995).

*Fundy basin—type area* The TS I is definitively present only in the Fundy basin, where it is known from two isolated sets of outcrops along the Bay of Fundy at Lower Economy, Nova Scotia, and in the vicinity of St. Martins, New Brunswick (Figure 10). Outcrops consist of alternating mostly fluvial and aeolian dune sequences and minor red lacustrine mudstone (Hubert & Mertz 1984, Nadon & Middleton 1985) (Figure 11). In Nova Scotia these comprise the Economy beds of the Wolfville Formation, and in New Brunswick they make up the Honeycomb Point Formation (Nadon 1981, Nadon & Middleton 1985). These strata, of Anisian age (Baird 1986, Olsen & Baird 1986, Olsen

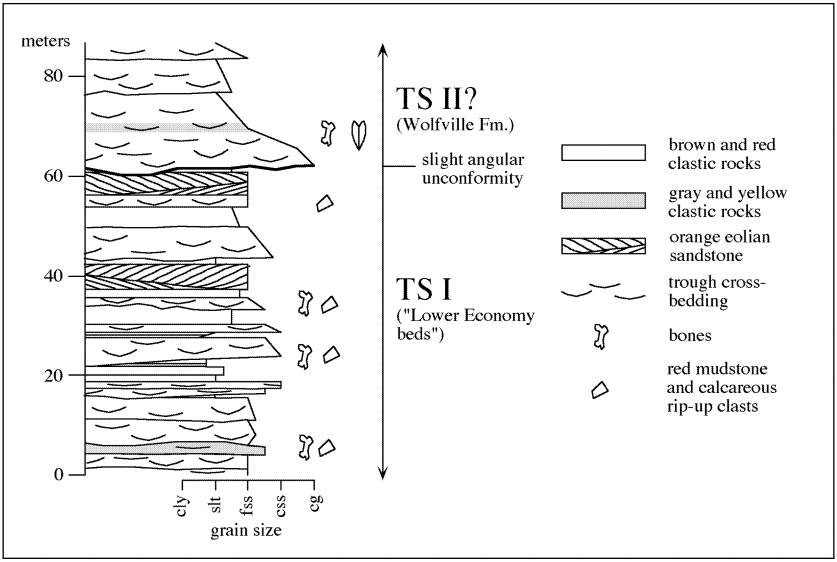


*Figure 11* Measured section of "Lower Economy beds" of TS I and unconformably (?) overlying Wolfville Formation of TS II at Lower Economy, Nova Scotia, Canada.

1988, Olsen et al 1989), thus comprise the oldest dated strata in the Newark Supergroup. Both the basal unconformity and the upper unconformity are exposed in the St. Martins area, but only the upper unconformity is exposed in Nova Scotia (Figure 11).

*Other basins* Sequences lithologically similar to those of TS I of the Fundy basin are present in the Argana Basin of Morocco, where they are termed the Ikakern Formation of Duffaud et al (1966) (t1 and t2 of Tixeront 1973) (Figures 3, 10). There is strong disagreement on the age of these strata, with opinions ranging from Permian (Dutuit 1988) to Anisian (see Medina 1991); hence the identity of the Ikakern as part of TS I is uncertain. However, there is little doubt of the unconformity separating the Ikakern of the overlying strata of unquestionable Triassic age, as shown in the figures of Tixeront (1973) (Figure 10). The TS I may also be represented in Morocco by middle Triassic–age fine red clastics, dolomites, and evaporites at the base of the Triassic sequence, found from drill holes in the Essaouira basin of the continental shelf of Morocco (Beauchamp 1988). The TS I has not been recognized in the Newark Supergroup.

TECTONOSTRATIGRAPHIC SEQUENCE II The TS II consists of early synrift sedimentary rocks comprised of strongly wedge-shaped sequences in basins larger than those of TS I, with lacustrine rocks that tend to be dominant in the basin's center. The TS II sequences exhibit evidence of deposition in more humid environments than those of TS I, as well as those of the younger sequences in the same basins. They are separated by a gentle angular unconformity from younger rocks in the up-dip portions of the fault blocks they comprise (Figure 9). The age of the sequence is Ladinian (?) to early Late Carnian [Middle (?) to early Late Triassic]. The TS II is recognized in the Richmond, Taylorsville, Farmville, and associated basins of Virginia (Goodwin et al 1986); the Fundy basin of Maritime Canada; and at least the Argana basin of Morocco. It may be much more widespread. The best-documented and type area is in the Richmond and Taylorsville basins of Virginia.

*Richmond and Taylorsville basins—type area* In the Richmond basin of Virginia, TS II comprises most of the sequence preserved in the basin. Cornet (in Cornet & Olsen 1990) recognized that the older sequence consists of the long-established (Shaler & Woodworth 1899) Tuckahoe Formation (originally group) that is made up of the "lower barren beds," "productive coal measures," and Vinita beds. However, the younger sequence consists of a previously unrecognized lacustrine interval, which Cornet named the Turkey Branch Formation, that is in turn overlain by Shaler & Woodworth's (1899) Otterdale sandstone. This younger sequence belongs in TS III (see below). The presence of abundant *Aratrisporites* in the lower part of TS II suggested to Ediger (1986) a Ladinian age, but Cornet & Olsen (1990) believe the age of these beds to be Early Carnian (Julian). The upper portion of the Tuckahoe Formation could be as young as early Late Carnian (early Tuvalian).

TS II, in the Richmond basin, is an example of a tripartite lacustrine sequence. The basal, so-called lower barren beds consist of a thin veneer of fluvial gray sandstone and conglomerate resting on basement. This is succeeded by the "productive coal measures" that consist of alternating gray sandstones, thin bedded and rooted gray and black mudstones, and coal seams that are locally up to several meters thick (Robbins et al 1988, Wilkes 1988). The overlying Vinita beds are a cyclical sequence of black and gray laminated mudstones and gray sandstones (Figure 12), made up of muted double–Van Houten cycles, with two high-stand intervals (e.g. Figure 5). The lower high-stand part of these cycles is thicker, more finely laminated, and more organic rich than the upper high-stand part (Ediger 1986). These cycles are similar to the Van Houten cycles seen in TS III, described below for the Dan River Basin (Figure 5). However, in contrast to the cyclicity seen in younger Triassic or Jurassic rift basin deposits in eastern North America, the cycles usually show little or no evidence of desiccation or subaerial exposure. Ediger ascribed these double cycles to the $\sim$40-ky cycle of the Earth's axial obliquity, modulating the $\sim$20-ky cycle of

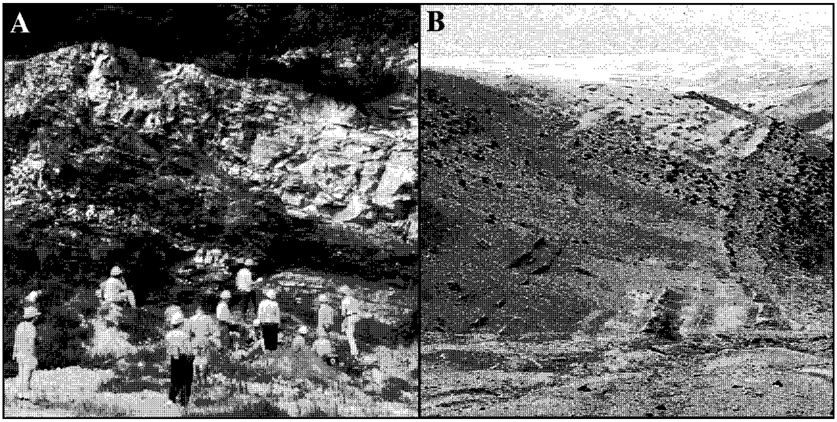


*Figure 12* Outcrops of TS II in the Richmond basin (A) and the Argana basin (B). (A) Outcrop of well-developed lacustrine black mudstones and gray sandstones of double Van Houten cycle in Vinita Beds in the Boscobel Quarry, Manakin, Virginia, United States. Van Houten cycle extends from head of person on far left to top of thick light-colored (*gray*) bed of middle right. There are no signs of subaerial exposure in this outcrop. (B) Distance view, looking north, of lacustrine Aglegal Mb. (t4 of Tixeront 1973) showing cyclical pattern of gray carbonates (*thin white bands*), purple and red mudstones (*light and darker gray broad bands*), and brown sandstones (*prominent benches*). The outcrops are near Tizi Maachou, Morocco. Section dips to left and thickness of prominent bench on right is about 2 m.

climatic precession. Although there is no direct evidence of the duration of these cycles, the theoretical considerations of the effects of celestial mechanics on insolation (Crowley et al 1992), and comparison with the double–Van Houten cycles in TS III, suggest that the Vinita cycle doublets are in fact a doubling of the frequency of the cycle of climatic precession itself as expected at the equator.

In the Vinita beds, deposition of the microlaminated units was probably in deep perennially chemically stratified water, whereas that of the less well-laminated units in the cycles was under shallower and higher energy water. The coals were deposited in swamps early in the depositional history of the TS II. Based on the presence of coals and thick black shales, limited signs of subaerial exposure, and a complete lack of evaporites, conditions seem to have been generally very wet.

The Taylorsville basin of Virginia and Maryland is located a few kilometers north of the Richmond basin (Figures 1, 7), to which it was most likely connected during deposition. The lower units of the Taylorsville basin (Doswell Formation; Figure 3, Table 1) belong to TS II, and are very similar to those of the Richmond basin (Weems 1980, Cornet & Olsen 1990, Olsen et al 1982, LeTourneau 1996), although no minable coals have been recognized. Over 6 km of continuous core has been recovered from the Taylorsville basin and is now under study (LeTourneau 1996).

*Other basins* In the Argana basin of Morocco, TS II is a tripartite sequence that makes up the Timezgadiwine Formation of Duffaud et al (1966). This formation consists of a basal conglomerate and sandstone (t3 of Tixeront 1973) overlying the unconformity with the Ikakern Formation, followed by cyclical pink and gray, locally carbonate-rich mudstone (t4 of Tixeront 1973) that grades up into cyclical red mudstone and sandstone (t5 of Tixeront 1973) (Figures 12, 13). The cyclicity, although evident in outcrops, has not been described. The lower part of the formation appears to be Early Carnian (Julian) and the upper part is late early Late Carnian (early Tuvalian) (Hunt & Lucas 1991). Based on Carnian-age palynomorph assemblages (Cousminer & Manspeizer 1976, Biron & Courtinat 1982), the Oukaimeden Sandstone of the High Atlas area of Morocco may also belong in TS II.

In the Fundy basin of the Maritime Provinces of Canada, TS I is unconformably overlain by TS II, which otherwise rests on prerift basement. In outcrops, TS II consists of fluvial strata of the Wolfville Formation (Powers 1916, Klein 1962) in Nova Scotia, and the Quaco and Echo Cover formations in New Brunswick (Magnusson 1955). The eolian strata and minor interbedded lacustrine rocks usually grouped with the Wolfville Formation (Hubert & Mertz 1984) actually belong in TS III (see below). The main mass of the Wolfville Formation consists of large (2–10 m) fining-upward cycles that were deposited by braided rivers, are extensively bioturbated, and have minor amounts of caliche (Hubert & Forlenza 1988, Smoot 1991). In the St. Martins area of New Brunswick, Canada, the Quaco and Echo Cove formations overlie the Honeycomb Point Formation of TS I in a well-exposed gentle angular unconformity (Figure 10) (Nadon & Middleton 1984). The Quaco Formation is a thick (190–300 m) clast-supported well-rounded conglomerate deposited in a braided (and probably anastomosing) river in a humid distal fan setting (Nadon & Middleton 1984). The overlying Echo Cove Formation consists of gray and pale-red pebbly sandstones and mudstones with locally abundant plant remains (Nadon & Middleton 1985). Based on the presence of conglomerates and bed forms that require relatively sustained flow, the dominant gray color and abundant plant remains of the Echo Cove Formation, and the absence of aeolianites, deposition of the Wolfville, Quaco, and Echo Cove formations were deposited under more persistently humid conditions than the underlying units of TS I. The age of TS II in the Fundy basin is Julian (Early Carnian) or Ladinian to early Late Carnian (early Tuvalian) age (Nadon & Middleton 1985, Hunt 1993; SJ Fowell, personal communication).

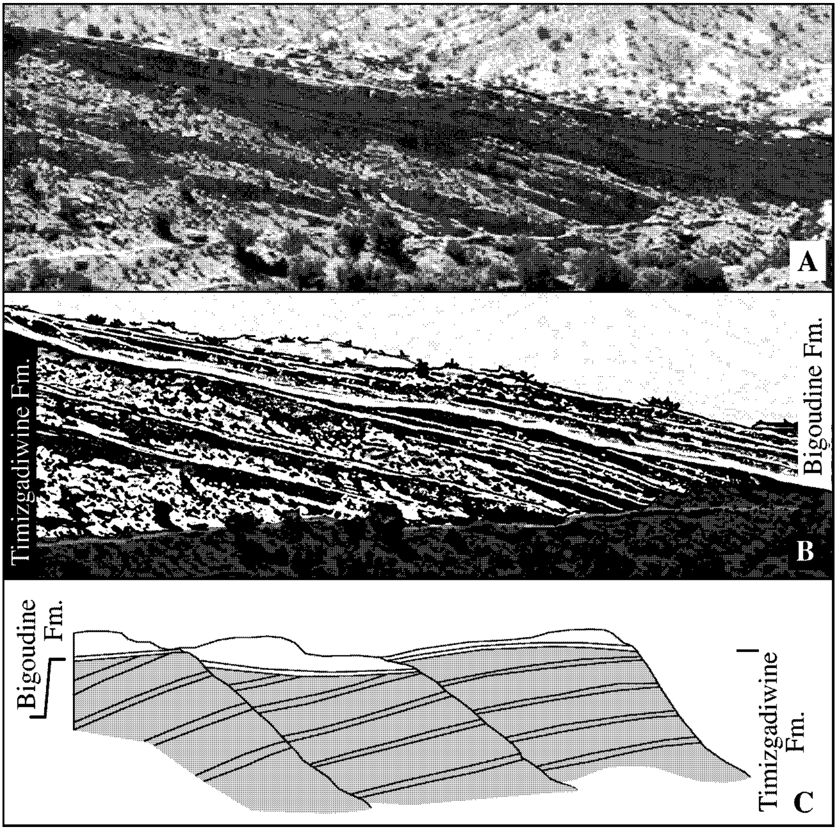


*Figure 13* Outcrops of unconformity between Timezgadiwine Formation (TS II) and overlying Bigoudine Formation (TS III) in the Argana basin. (A) Photograph of outcrop of lacustrine red mudstones and sandstones of the Timezgadiwine Formation overlain by brown fluvial conglomerate and eolian sandstone of Bigoudine Formation east of Argana, Morocco. (B) Same as (A) but enhanced to show bedding and position of angular unconformity. (C) Sketch of outcrop from Argana basin, redrawn from Tixeront (1973).

Unequivocal lacustrine strata have not been identified in the two industry exploratory drill holes that penetrate the Wolfville Formation or in the outcrops of the Quaco and Echo Cove formations (Olsen et al 1989). However, seismic profiles from the Chignecto subbasin of the Fundy basin show a sequence low in the basin section (Figure 7) characterized by laterally extensive high-amplitude reflections (Withjack et al 1995). This package fans towards the border fault and is separated by a gentle unconformity from overlying less-reflective units. Based on the reflection character and geometry, it may be a lacustrine equivalent of the Wolfville Formation, as has been postulated to exist by Brown (1986). However, until the unit is drilled, its character will remain uncertain.

Other units that probably belong in TS II (see Figure 3) include the following: (*a*) portions of the distinctive "Millstone grit" of the Deep River basin of North Carolina (Stagg et al 1985), based on pollen and spores (Cornet 1977, Litwin & Ash 1993); (*b*) basal beds of the Gettysburg basin of Pennsylvania (Cloos & Pettijohn 1973, Stose & Jonas 1939, Cornet 1977, Smoot 1991a), based on age and facies; (*c*) the basal units of the Newark basin of New York, New Jersey, and Pennsylvania, based on seismic profile character (Figure 7) (Costain & Coruh 1989); (*d*) basal units in the Norfolk basin of the Coastal Plain continental shelf of Virginia, again based on seismic character (Figure 7) (Musser 1993); and (*e*) the Kettle red beds at the base of several basins on the Scotian Shelf and Grand Banks of Newfoundland, based on stratigraphic position and scant age data (Barss et al 1979).

TECTONOSTRATIGRAPHIC SEQUENCE III Middle synrift sedimentary rocks of TS III form the bulk of the basin fill throughout eastern North America and Morocco, and they are also geographically the most widespread units (Figures 2, 3). In general, the sequence thickens less toward the border faults than any of the other tectonostratigraphic sequences. The age of the strata is Late Carnian to late Rhaetian (early Late Triassic to latest Late Triassic). Thicknesses can exceed 6 km (Olsen et al 1986a). There is often a thick (>500 m) basal, fluvial interval followed by a thicker (>1 km) lacustrine sequence. In some basins the middle synrift sedimentary rocks are entirely fluvial. These are the "typical" red beds of northern Pangea. On the Scotian Shelf, the Grand Banks, and possibly Georges Bank, extensive Triassic-Jurassic evaporites are found in this sequence. The best-known portion of TS III is the Newark basin of New York, New Jersey, and Pennsylvania, which is considered the type area of the sequence (Figures 2, 6, 7, 15).

*Newark basin—type area* Three formations comprise TS III in the Newark basin. These are the Stockton Formation of fluvial-deltaic origin, the Lockatong Formation deposited mostly in lakes, and the Passaic Formation of lacustrine and fluvial origin. All of the Newark basin section except the lowest Stockton Formation and uppermost Boonton Formation have been continuously cored by the Newark Basin Coring Project (NBCP), and thus TS III in the Newark basin is the best-known stratigraphic section of any Triassic-Jurassic rift (Figure 6).

The Stockton Formation consists of more than 900 m (NBCP composite thickness; Olsen et al 1996a) of Carnian age gray, purple, and buff sandstones and red and gray mudstones. The upper half is arranged in a large-scale cyclical sequence mirroring the overall pattern of the overlying Lockatong Formation. Turner-Peterson (1980), Turner-Peterson & Smoot (1985), and Smoot (1991a) have described the upper Stockton Formation as deposited predominantly by meandering rivers with deltaic and lacustrine components (see also Reynolds 1993). According to Smoot (1991a), the lower Stockton Formation was deposited primarily by braided rivers. Based on industry drilling and seismic profiles, the Stockton Formation does not change laterally into fully lacustrine mudstones within the basin (Reynolds 1993) but rather retains its basic character to very near the border fault zone, where it passes into coarse conglomerates (Allen 1979, Olsen 1980a).

The Lockatong and overlying Passaic formations comprise an enormous (>5 km) predominately lacustrine package that is mostly gray and black in the Lockatong Formation and mostly red in the Passaic Formation. Together the formations span from the Late Carnian (late Tuvalian) to the earliest Hettangian. The entire thickness is characterized by a pervasive hierarchy of sedimentary cycles (Figures 4, 6, 14) (McLaughlin 1933, Van Houten 1962, 1964, 1969, 1980, Olsen 1980c, 1986, Smoot & Olsen 1985, 1988, 1994, Smoot 1991a, Olsen et al 1996a, Olsen & Kent 1996). In the more central part of the basin the Lockatong and Passaic Formations are largely mudstone, but along the edge they grade into coarser clastic rocks, which also represent a progressively larger proportion of area upward through the two formations. Evaporites, apparently mostly sulfates or pseudomorphs after sulfates, are fairly abundant in the fine-grained facies of the upper Lockatong through middle Passaic Formations (Smoot & Olsen 1994, El-Tabakh 1994, El-Tabakh & Schreiber 1994, El-Tabakh et al 1996, Olsen et al 1996a, Riccioni et al 1996). Very thin eolian sands occur sporadically in the upper Passaic Formation (Smoot & Olsen 1994). The uppermost Passaic Formation members, the Pine Ridge Member and Exeter Member, probably belong in TS IV (Figure 16, see below).

Spectral analysis of the NBCP cores shows a very strong dominance of the ∼20-ky cycle of climatic precession and its modulating "eccentricity" cycles of ∼100-ky, ∼413 ky, and ∼2 my (Figures 4, 5). There are hints of the double (i.e. ∼10 ky) low in the Lockatong Formation, when the Newark basin was close to the paleoequator (about 4°N). Black laminated shales tend to comprise the largest proportion of the cycles in the lower Lockatong Formation, and this proportion rhythmically decreases upward (Figure 14).

Newark basin Van Houten cycles in TS III record the deepening and shallowing of lakes. At one end of the range are cycles with microlaminated organic-rich deep-water deposits formed by lakes that developed into very deep (+80-m depth) perennially chemically stratified lakes that shallowed to playas. At the other extreme are Van Houten cycles with a relatively deep water portion represented only by red thin bedded mudstone with widely spaced desiccation cracks. These were produced by playas that had a protracted filling episode, shallowing to evaporite-producing pans or even vegetated flats. The various

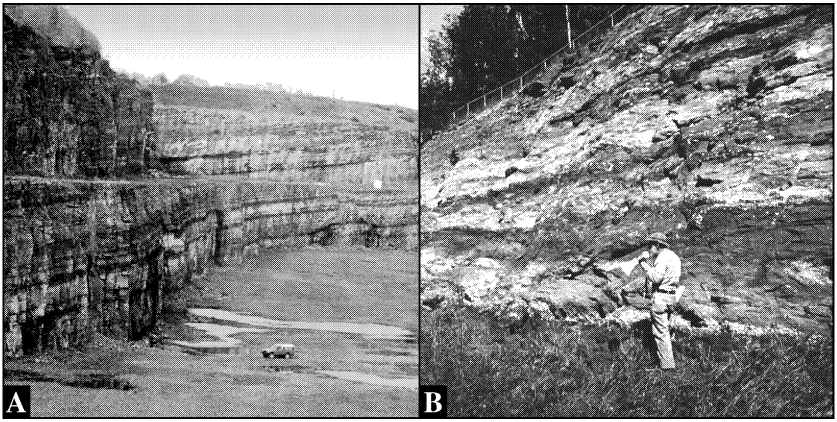


*Figure 14* Exposures of TS III in the Newark basin (A) and the Hartford basin (B). (A) Lacustrine lower Lockatong Formation in quarry in Eureka, Pennsylvania, United States. Prominent black shales mark out the deepest-water portions of Van Houten cycles; note pickup truck for scale. (B) Braided stream sandstones (*light*) and mudstones (*dark*) of New Haven Formation. Prominent nodular caliche beds are just above the grass line on right and above and right of person's (JF Hubert) head.

types of Van Houten cycles, their interpretation, and the details of the vertical changes in type are described by Smoot & Olsen (1994).

Laterally, the lower part of the Lockatong Formation passes into mostly gray and buff deltaic and fluvial sandstones (Olsen 1980b, Parker et al 1988), and the upper part of the formation becomes dominated by red clastics and then merges with the Passaic Formation (Olsen 1980b). Both the latter and the Passaic Formation in general grade laterally into red coarse clastics approaching the edges of the basin, although in many cases, some of the better developed black shales within individual cycles can be traced into conglomerates to the basin edge (Olsen et al 1989). Fluvial clastics dominate in the change toward the edges along the basin axis, but alluvial fan strata dominate toward the border fault system. These coarse clastics have been described by Picard & High (1963), Glaeser (1966), Arguden & Rodolfo (1985), Van Houten (1969), Ratcliffe (1980), Ratcliffe et al (1986), Olsen et al (1989), Parker et al (1988), and Smoot (1991a).

Glaeser (1966) has shown that the provenance of most of the Stockton and Lockatong formations was from the southeast (the hanging wall), whereas for much of the Passaic Formation most sediments were axially fed (Smoot 1991a), probably mostly from footwall rocks. This change in source may reflect a decreasing magnitude of basin subsidence and hence footwall uplift throughout the evolution of TS III in the Newark basin. These changes in provenance are reflected in paleocurrent patterns as well (Parker et al 1988, Smoot 1991a).

Based on the NB-1 line (Figure 7), a line taken by Exxon in the southwestern part of the basin (Reynolds 1993), and the NBCP cores (Olsen et al 1996a, Schlische 1992), the overall geometry of TS III is that of a gently tapering wedge, slowly fanning toward the border fault system. There are also some indications that the major internal faults of the basin were active during the deposition of TS III, which resulted in minor thickening of strata toward those faults as well.

*Other basins* Going south from the Newark basins, the general trend is toward increased wetness, both in general climate-sensitive facies and in cyclicity—at least in the exposed basins (Figure 3). The closest basins to the south of the Newark basin are the Gettysburg and Culpeper basins. The former is physically continuous with the Newark basin, and the latter basin is separated from the Gettysburg basin by only a very narrow strip of basement. Thus, during deposition, they were subbasins of a very large rift basin that was comparable in size to the present-day Baikal or Tanganyika rifts. As might be expected, there is considerable continuity of facies between these basins in TS III (Figure 15).

The stratigraphy of the Gettysburg basin of Pennsylvania and Maryland remains poorly known (Stose & Bascom 1929, Stose & Jonas 1939, Cornet 1977, Smoot 1991a). Apart from the absence of a Lockatong Formation–like facies, which is replaced by a fluvial-lacustrine Stockton Formation–like unit in the middle and upper New Oxford Formation (Olsen et al 1982), the section is similar to that in the Newark basin. The overlying Gettysburg Formation is lithologically very similar to the Passaic Formation and shares a very similar cyclical pattern (Smoot 1991a), although quantitative studies have not been carried out. Evaporites are not common as in the Passaic Formation, and they are absent from the New Oxford Formation.

Smoot (1991a) has stressed the similarities between the Culpeper and Gettysburg sections. In the Culpeper basin, a lower part (Manassas Formation) is similar to the New Oxford Formation of the Gettysburg basin and is overlain by a series of formations (Balls Bluff, Tibbstown, Catharpin Creek formations) that, together, are very similar to the Passaic Formation (Smoot 1991a). This sequence was described in broad outline by Roberts (1928) and in much more detail by Lindholm (1979), Lee (1979), Lee & Froelich (1989), Froelich et al (1982), Gore (1986, 1988a,b, 1994), Smoot & Olsen (1988), Olsen et al (1989), and Drake et al (1994). Preliminary magnetostratigraphic correlation of cores and outcrops of the Balls Bluff Formation from the southern Culpeper basin (Andrus no. 1 and Lenn Brothers no. 1 cores) with the NBCP cores of the middle Passaic Formation show an extremely close match in cyclostratigraphy (DV Kent, PE Olsen, PM LeTourneau, unpublished data). Based on Fourier analysis, the amplitude of the cyclicity (and general wetness of the facies) is substantially higher in the Culpeper than in the Newark basin, and, as in the Newark basin, the ~20-ky cycle of climatic precession and its modulating cycles dominate (Figure 5).

The entire preserved section of the Dan River–Danville basin appears to be within TS III, which correlates with the Stockton, Lockatong, and lower Passaic formations of the Newark basin (Olsen et al 1978, Robbins et al 1988, Litwin & Ash 1993, Olsen et al 1991). The magnetic polarity stratigraphy of this basin has been worked out by Kent & Olsen (1996), which allows for an independent detailed correlation. As understood from the new magnetostratigraphic work and previous mapping and stratigraphic work by others (Meyertons 1963, Thayer 1970, Robbins 1982, Thayer & Robbins 1992, 1994), the basin sequence consists of two sequential tripartite lacustrine sequences (Figure 3), apparently not separated from each other by an unconformity. The lower tripartite interval correlates with the upper Stockton Formation of the Newark basin (Kent & Olsen 1996). It has thin coal beds in cyclical gray and black mudstones and sandstones in its lower part and mostly red gray sandstones in its upper part. The succeeding tripartite sequence is very thick (~1.8 km), consisting of gray and black cyclical lacustrine mudstones that correlate with the Lockatong and basal Passaic formations of the Newark basin. These strata produce a rich insect and vertebrate assemblage (Olsen et al 1978, Fraser et al 1996). The youngest strata are cyclical red and gray clastics that correlate with the lower to middle Passaic Formation. Although the chronostratigraphy of the basin is known in some detail, the stratigraphic nomenclature is unnecessarily cumbersome owing to separate terminology in Virginia (Meyertons 1963) and North Carolina (Thayer 1970).

Fourier analysis, with a time scale provided from magnetic polarity stratigraphy, shows that the cyclicity in the gray and black portion of the upper tripartite sequence (upper member of the Cow Branch Formation of Olsen et al 1978) is dominated by the double precession cycle of ~10 ky (Kent & Olsen 1996). The Van Houten cycles tend to be doubled, with two significant black shale intervals (Figure 5). Rare pseudomorphs are the only indication of evaporites. Visual examination of the lower coal-bearing gray and black sequence shows that it too is dominated by the double Van Houten cycles. This pattern of cyclicity is consistent with the expectations of equatorial Milankovitch-type control of climate.

Depositional environments for the cyclical parts of the Dan River–Danville basin sequence have been described (Olsen et al 1978, Olsen 1983, Robbins 1982, Olsen et al 1989, Olsen & Johansson 1994, Fraser et al 1996). These cycles were deposited by lakes that fluctuated between deep perennially chemically stratified (meromictic) lakes and shallow ephemeral (but not evaporitic)

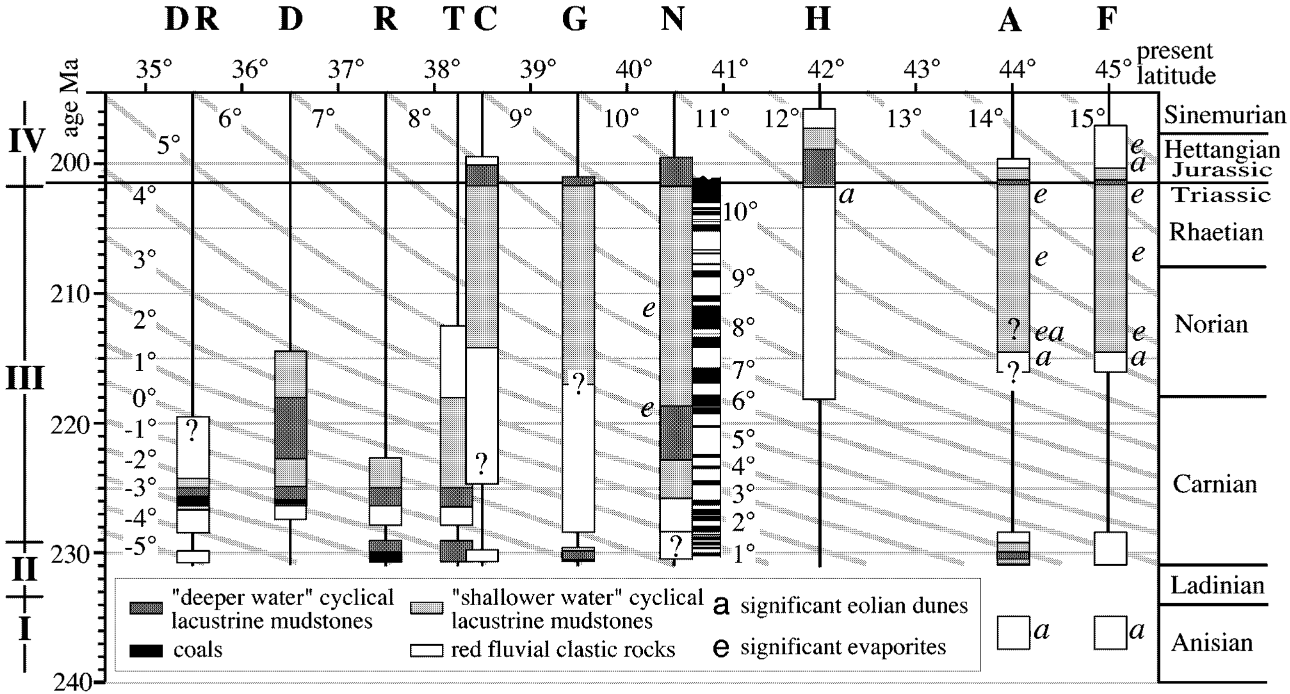


*Figure 15* Relationship between main climate-sensitive lithologies, age, and geography for the major Newark Supergroup basins and Morocco. Curved gray lines are lines of equal paleolatitude based on the Newark basin section (Newark basin polarity scale from Kent et al 1995). The tectonostratigraphic sequences of the CAM basins are indicated by I, II, III, and IV. Basins are as follows: DR, Deep River; D, Dan River–Danville; R, Richmond; T, Taylorsville; C, Culpeper; G, Gettysburg; N, Newark; H, Hartford; A, Argana; F, Fundy.

lakes and vegetated flats. The fluvial and marginal lacustrine environments (Thayer 1985, Smoot 1991a, Thayer & Robbins 1992) have been much less studied.

The TS III in the Richmond basin of Virginia (Turkey Branch and Otterdale formations) and the Taylorsville basin of Virginia and Maryland (formations not named) consists of a lower red and gray apparently fluvial or fluvial-lacustrine unit, which grades up into a cyclical mostly gray and black clastic unit. This is overlain by a red and gray marginal lacustrine to fluvial sequence. This sequence correlates with the upper Stockton through lower Passaic formations of the Newark basin (Late Carnian–middle Norian) (Cornet & Olsen 1990, Fowell 1993, LeTourneau 1996). Like correlative strata in the Dan River–Danville basin, the gray and black unit has a strong double Van Houten cycle as seen in cores and outcrops. There are no evaporites.

The Deep River basin of North and South Carolina consists of three subbasins with somewhat dissimilar stratigraphies, which apparently comprise strata mostly belonging in TS III based on paleontological data (Cornet 1977, Olsen 1977, Olsen et al 1982, Robbins et al 1988, Olsen et al 1991, Litwin & Ash 1993). Best known is the sequence in the Sanford subbasin (Reinemund 1955, Textoris & Gore 1994). In outcrop and core samples, the sequence consists of a tripartite sequence of a lower mostly red fluvial unit (Pekin Formation), succeeded by a lacustrine gray and black coal-bearing sequence (Cumnock Formation), which in turn is overlain by a lacustrine to fluvial red sequence (Sanford Formation). The Cumnock Formation is cyclical (Hu & Textoris 1994) and appears to consist of double Van Houten cycles of a muted aspect similar to that of the upper Vinita Formation of TS II (PE Olsen, unpublished data). Textoris & Gore (1994) report a gypsum-bearing playa sequence in the Pekin Formation based on drill cuttings, the presence of which I have not been able to confirm by subsequent examination of the cuttings. Such a deposit is very out of character with the quite humid aspect of outcrops of the Pekin Formation (Olsen et al 1991) or for that matter any part of the sequence in the Sanford subbasin. The meaning of these samples awaits further study.

The stratigraphic relationships of the TS III sections in the Wadesboro basin (Randazzo & Copeland 1976, Randazzo et al 1970, Cornet 1977, Olsen et al 1991) and the Durham basin (Wheeler & Textoris 1978, Textoris 1994, Gore et al 1989, Hoffman & Gallagher 1989, Olsen 1977, Olsen 1988, Olsen et al 1989, Huber et al 1993) to those of the Sanford basin remain relatively poorly defined. Gray to red lacustrine mudstones of apparently disparate ages are present in both subbasins.

Several basins, known from geophysics, lie below the Coastal Plain directly to the east of the Newark basin (Hutchinson & Klitgord 1988a,b, Benson 1992) (Figure 1), but none have been directly sampled or studied in detail. Basins in the subsurface are known in southern North America from Virginia to Florida (Gohn & Higgins 1978, McBride et al 1987, Costain & Coruh 1989, Manspeizer 1989, McBride 1991, Benson 1992, Musser 1993) and along the Gulf Coast to Texas (Manspeizer 1989, Bartok 1993). The Norfolk basin of eastern Virginia appears to have a well-developed tripartite sequence in TS III, based on seismic character studies (Musser 1993). Farther south, basins such as the Florence and Dunbarton basins beneath the South Carolina coastal plain consist of, where sampled, primarily red and gray massive mudstone, sandstone, and conglomerate sequences that apparently belong to TS III (Steele & Colquhoun 1995, Marine & Siple 1974, La Tour et al 1995). Sedimentary sequences of the South Georgia rift system of Georgia and Florida (Weaver 1976, Chowns 1985, Bowman 1985) and the Eagle Mills Formation (Burch 1992, Burch & Weidie 1994, Dawson & Callender 1992) of the subsurface Gulf coast also consist of red and gray clastic rocks, although available descriptions are sketchy. Very few indications of lacustrine strata are present in these subsurface southern basins. An exception may be the south Georgia rift where some cyclical red, gray, and black mudstones have been reported on the basis of well cuttings (Lightsey number 1 well; B Cornet, personal communication). The scant paleontologic, core, and seismic profile data suggest that these sequences pertain to TS III (Moy & Traverse 1986, Traverse 1986).

Going north of the Newark basin, TS III is represented by outcrop in the Pomperaug, Hartford, Deerfield, Argana (and related basins), Fundy, and probably Chedabucto basins. In the Pomperaug, Hartford, and Deerfield basins of Connecticut and Massachusetts, red and buff fluvial (braided stream) clastic rocks strongly predominate (Hobbs 1901, Hubert 1977, 1978, Hubert et al 1978, Stevens & Hubert 1980, Weddle & Hubert 1983, Smoot 1991a, McInerney 1993, Horne et al 1993) (Figure 16). TS III rests directly on crystalline basement and is bounded upward by either the lower lava flow sequence of TS IV or the first sedimentary formation of Jurassic age. Red fine-grained sequences in the Hartford (New Haven Formation) and Deerfield basins tend to have well-developed caliche profiles (Figure 14) (Hubert 1977, Schoelles 1995, Wang et al 1995), and local silcretes (McDonald & Textoris 1984) and meter-scale eolian dune sandstones occur (upper New Haven Formation) (Smoot 1991b). The environments represented by TS III in the Pomperaug, Hartford, and Deerfield basins appear drier than comparable facies in the Newark basin, and there are no lacustrine strata.

The TS III is well represented in the Argana basin of Morocco by the Bigoudine Formation (t6, t7, t8 of Tixeront 1973) (Duffaud et al 1966) (Figure 13). The basal portion of the formation consists of a lower red conglomerate and sandstone and an upper buff eolian sandstone sequence (Tadrart Ouadou Mb.) with some interbedded halite (Tixeront 1973). Recent investigations (JP Smoot,

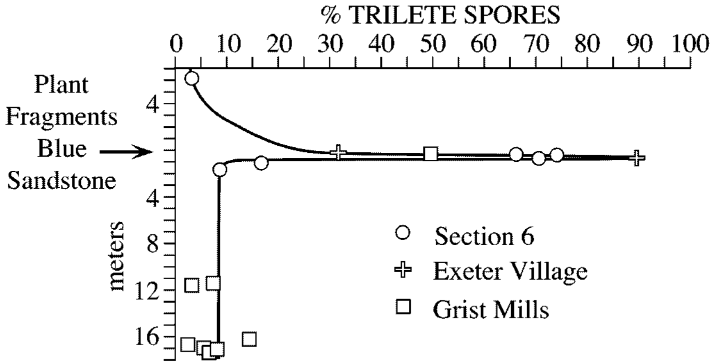


*Figure 16* Percentage fern spores around palynologically identified Triassic-Jurassic boundary in the Newark basin. Adapted from Fowell et al (1994).

PE Olsen, PM LeTourneau, unpublished data) agree with Tixeront (1973) in showing that the basal conglomerate rests unconformably on the Timezgadiwine Formation (TS II). Overlying the eolian beds are red-brown cyclically bedded sandstones and mudstones with a few cyclical gray and black shales and thin eolian dune sands (Sidi Mansour Mb.) of Norian age (Fowell et al 1996b), which are in turn overlain by cyclical red mudstones (Hasseine Mb.). The uppermost few meters of the latter are a series of centimeter-scale palyniferous gray and black mudstones that contain the Triassic-Jurassic boundary (Fowell et al 1996a) and probably belong in TS IV.

Smoot & Olsen (1988) have shown that both the Sidi Mansour and Hasseine Mbs. are composed largely of "sand patch cycles" produced in salt-encrusted saline mudflats. These consist of red mudstone beds passing upward into massive sandy mudstone or muddy sandstone that are comprised of irregular pods of sandstone and siltstone with a distinctive cuspate geometry. Evaporite molds as well as crystals and nodules of gypsum are abundant, and sand patch mudstone requires the presence of halite or other very soluble salts to form (Smoot 1991a). The gray and black shales are the high-stand portions of transgressive-regressive sequences that resemble Van Houten cycles within the Passaic Formation of the Newark basin. Note that these interpretations differ strongly from those of Brown (1980).

Thus, the Bigoudine Formation appears to have been deposited under an arid but fluctuating climate. Evaporitic pans were the predominant depositional system. The Bigoudine Formation stands in marked contrast to the underlying Timezgadiwine Formation, which was deposited under more humid conditions, and closely resembles TS III in the Fundy basin (below).

Other Triassic sequences in Morocco and on the Moroccan continental shelf have extensive deposits belonging to TS III. To the east, they tend to be more detrital ("upper siltstone" unit; Beauchamp 1988), but to the east and north they tend to contain thick halite sequences (Amadé 1965, Tortochaux 1968, Van Houten 1977, Salvan 1984, Holser et al 1988). In the far north of Morocco (Rif) they grade into marine carbonates. Strata in TS III on the Moroccan meseta tend to be relatively thin and, interestingly, are not syndepositionally tilted (Salvan 1975). According to Holser et al (1988) and Krueger & Reesman (1983), the evaporites in the Argana basin and Ourika area (Atlas province of Holser et al 1988) were deposited in local continental playas or sabkas fed by nonmarine waters. In contrast, those of the eastern and northern areas (Atlantic province of Holser et al 1988) have salt deposits indicating an interconnected, probably marine source, with a significant nonmarine input.

In the Fundy basin, TS III is represented by the upper Wolfville Formation and the Blomidon Formation and in total is very similar to the Bigoudine Formation of the Argana basin (Smoot & Olsen 1988). The upper Wolfville Formation consists of a lower fluvial sequence with intercalated and overlying minor mudstones and eolian dune sands, some of which are locally quite thick (Hubert & Mertz 1984, Olsen et al 1989). The Blomidon Formation overlies the upper Wolfville Formation and consists mostly of sand patch cycles (Smoot & Olsen 1985, 1988; Olsen et al 1989). Descriptions and interpretations of the Blomidon Formation have been given by Klein (1962), Hubert & Hyde (1982), Olsen et al (1989), and Mertz & Hubert (1990). There are abundant nodules and crystals of gypsum and occasional beds of eolian dune sand (Smoot & Olsen 1988). The middle of the formation contains a sequence of interstratal karst deposits, which are probably dissolution sequences after halite (Olsen et al 1989, Ackermann et al 1995) (Figure 17), associated with fish-bearing laminated red, yellow, and purple mudstones (Carroll et al 1972). The upper few meters of the Blomidon Formation contain a sequence of thin gray and black palyniferous or plant-bearing mudstones (Carroll et al 1972, Fowell & Traverse 1995) very similar to those of the Argana basin, which probably belong in TS IV. The thin-bedded and laminated red, purple, gray, and rarely black mudstone beds within this overall sequence of sand patch cycle are laterally continuous at the scale of 60 km, based on outcrops and core samples (taken by Getty Mines in 1977; Cameron & Jong 1985). Together, these mudstones and the sand patch cycles make up a hierarchy of higher order cycles suggestive of Milankovitch origin.

Preliminary magnetostratigraphic correlation with the Newark basin section suggests that the entire Blomidon Formation correlates with the middle and upper Passaic Formation. This correlation corresponds to an average sedimentation rate of about 0.03 mm/yr, about an order of magnitude less than the Newark basin (DV Kent & PE Olsen, unpublished data). It also suggests that

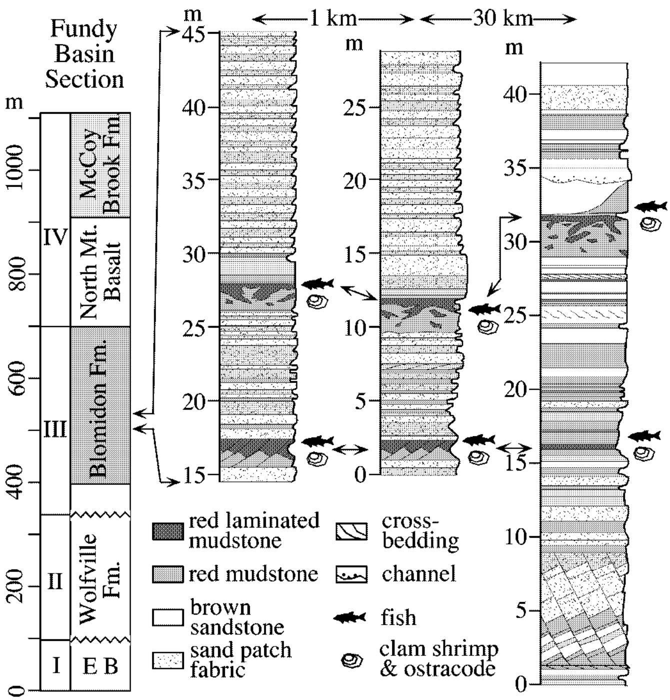


*Figure 17* Sand patch cycles and salt collapse intervals in three correlative outcrops along shore outcrops of geographic Minas basin, Fundy rift basin, Nova Scotia, Canada. The most prominent collapse features are (from *left* to *right*) between 25 and 30 m, 7 and 13 m, and 23 and 35 m in the three sections. Redrawn from Ackermann et al (1995).

the sand patch cycles (~1.3 m/cycle) had a duration of about 40 ky, which indicates an origin from the obliquity cycle rather than the climatic precession cycle that is dominant farther south. Longer frequency cycles are also present and preliminary work demonstrates cycles with periods of ~100 ky, ~400 ky, 1 my, 2 my, and 4 my, which suggests the modulating cycles of both climatic precession and the obliquity cycle.

Little is known about the many early Mesozoic strata inferred to exist on the continental shelf off eastern North America, south of the Scotian Shelf (Figure 1). Compilations of their distribution have been provided by Olsen (1977), Manspeizer (1988), Manspeizer & Cousminer (1988), Hutchinson & Klitgord (1988a,b), Klitgord et al (1988), and Benson (1992). The only well that plausibly encountered Triassic strata in this large region was the COST G-2 in the Georges Bank basin, on the continental shelf off Massachusetts (Schlee

& Klitgord 1988); however, its stratigraphy is very controversial. Fairchild & Gartner (1977), Bebout (1980), and Poag (1982) correlated salts near the base of the well with the Early Jurassic Argo Formation salts of the Scotian Shelf and the Grand Banks, but Cousminer et al (1984) and Cousminer & Steinkraus (1988) give a Late Triassic (Carnian-Norian) age. The differences remain unresolved (Manspeizer 1988, Poppe & Poag 1993) but are important, because the thick salt sequence would indicate that a Triassic incursion of presumably marine waters into an at least occasionally arid environment occurred farther south than previously recognized. However, the position of the well is not much farther south than predrift positions of the Moroccan Triassic age salts.

The successful search for hydrocarbons on the Scotian Shelf and on the Grand Banks of Newfoundland has resulted in much more knowledge of the numerous Triassic-Jurassic basins there than in subsurface basins to the south. More recent reviews of the stratigraphy of these basins include Holser et al (1988), Welsink & Tankard (1988), Tankard & Welsink (1988), Austin et al (1989), Welsink et al (1989), Tankard et al (1989), Jansa et al (1980), McAlpine (1990), and Pe-Piper (1992). In most of the drilled early Mesozoic–age sections there are extensive salt sequences interbedded with mostly red clastic rocks. The older sequence of red beds and salt (Osprey Evaporites) occur in various basins on the Grand Banks (e.g. Carson and Jeanne d'Arc basins; Jansa et al 1977, 1980), and are of Carnian-Norian age. Younger red beds and salts of Norian to Early Jurassic age (Eurydice Formation and Argo Salt) occur in basins both on the Scotian Shelf and on the Grand Banks (Holser et al 1988). In some cases over 1.4 km of Triassic-Jurassic salt has been drilled. These thicknesses may be exaggerated by salt tectonics, however (MO Withjack, personal communication). In wells on the Scotian Shelf (Glooscap C-63) and on the Grand Banks (Spoonbill C-30, Cormorant N-83; Pe-Piper 1992), tholeiitic basalt flows have been encountered and thus place an upper limit on TS III. In the Glooscap C-63, the upper part of TS III consists of salt of the Argo Formation succeeded by red shale to the base of the basalt. In the Spoonbill C-30 well, salt of the Osprey Formation is succeeded upward by red beds of the Eurydice Formation and finally dolomite at the base of the basalt; in the Cormorant N-83 well the sequence is similar, except that no salt was encountered below the basalt. The Triassic-Jurassic boundary was placed by Williams (1975) at the base of the dolomite in the Cormorant N-83 well, and the top of TS III is probably not far beneath it. In most places, however, there is no basalt, and the Triassic-Jurassic boundary and the base of TS III is within either the red beds of the Eurydice Formation or in the Argo Salt. These sequences bear a marked similarity to the exposed and buried Moroccan examples of TS III (Koning 1993).

Outcrops of red clastic rocks along Chedabucto Bay are probably equivalent to the offshore Eurydice Formation of the Orpheus Graben (Lyngberg 1984).

The Chedabucto Formation was named by Klein (1962) and has produced fragmentary remains of dinosaurs of probable Late Triassic age (Carroll et al 1972). The landward portion of this rift certainly represents TS III.

TECTONOSTRATIGRAPHIC SEQUENCE IV Late synrift rocks, usually beginning with an extensive sequence of tholeiitic lava flows and interbedded sedimentary strata, overlie TS III in many CAM basins from the Culpeper basin north to the Grand Banks and Morocco (Figures 1, 18). Sediments interbedded and overlying the basalts typically have much higher sedimentation rates than underlying sequences and, in exposed basins, have much better developed lacustrine high-stand lake deposits, suggesting more humid depositional environments or greater lacustrine basin depth. Tholeiitic lava flows without extensive interbedded sediments may represent TS III in the southeastern United States. Present evidence suggests a very brief but massive episode of flood basalt extrusion, lasting perhaps 600 ky (Olsen et al 1996b), covered as much as 2.3 million km$^2$ (McHone & Puffer 1996). The age of TS IV is latest Triassic through Early Jurassic (Cornet 1977, Cornet & Ziegler 1985), although it may get as young as Pliensbachian (Cornet 1977, Olsen et al 1989). In the exposed basins, this sequence marks out another tripartite lacustrine sequence.

Radiometric dates from the lava flows themselves traditionally have been difficult to interpret because of extensive alteration (see Sutter 1988, Seidemann 1989). However, U-Pb dates from granophyric veins from the North Mountain Basalt of the Fundy basin and U-Pb and $^{40}Ar/^{39}Ar$ dates from a feeder intrusion of the Newark basin flows indicate dates of about 201–202 Ma for the flows (Hodych & Dunning 1992, Ratcliffe 1988, Sutter 1988). The Milankovitch cyclostratigraphy of the surrounding sedimentary formations in the Deerfield, Hartford, Newark, and Culpeper basins is the basis for the suggestion that the entire extrusion sequence lasted no longer than about 600 ky (Olsen et al 1996b). The duration of the CAM basin igneous episode is thus similar to that of other flood basalt provinces, such as the Deccan and Siberian traps (Jaeger et al 1989, Renne & Basu 1991).

The TS IV probably includes the uppermost sedimentary sequence underlying the oldest lava flows. These sedimentary units include the Triassic-Jurassic boundary. In places where there is high-resolution data (i.e. the Newark basin), no evidence exists for an unconformity between TS III and IV (Fowell & Olsen 1995). However, because half graben subside asymmetrically (Schlische 1992) and TS IV is characterized by accelerated accumulation, increased subsidence rates, and half graben asymmetry, up-dip unconformities might be expected and in fact may have been observed locally in the Hartford basin (see below).

*Hartford basin—type area* Tholeiitic lava flows and the interbedded and overlying sedimentary sequences in the Hartford basin (Figure 2) have been well

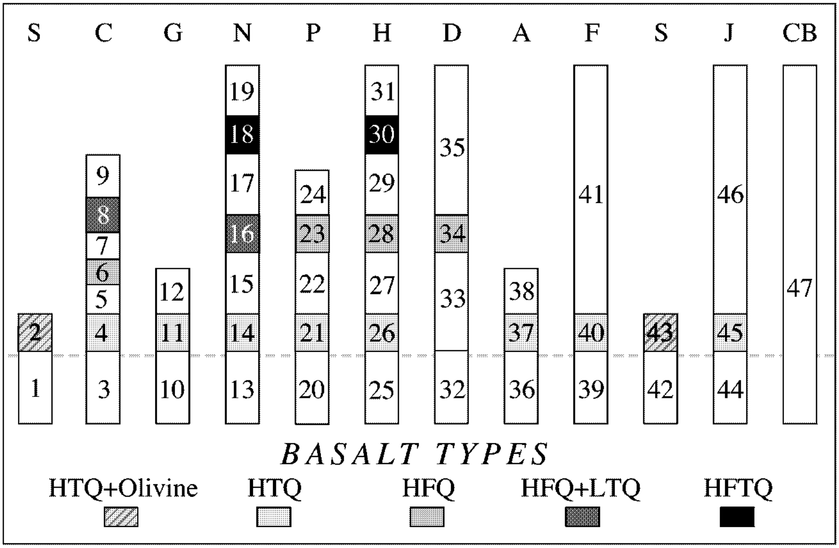


*Figure 18* Relative time and chemical relationships of basalts and sedimentary formations within TS IV. Basins are as follows: S, South Georgia basin (South Carolina, Georgia); C, Culpeper basin (Virginia, Massachussetts); G, Gettysburg basin (Pennsylvania); N, Newark basin (Pennsylvania, New Jersey, New York); P, Pomperaug basin (Huber & McDonald 1992) (Connecticutt); H, Hartford basin (Connecticutt, Massachussetts); D, Deerfield basin (Massachussetts); A, Argana basin (Morocco); F, Fundy basin (Maritime Canada); M, Mohican basin (Scotian Shelf, Nova Scotia, Canada) (Glooscap C-63; Pe-Piper et al 1994); J, Jeanne d'Arc basin (Grand Banks, Newfoundland, Canada) (Amoco-Imp, A-1 Cormorant N-83 well; Pe-Piper et al 1984); CB, Carson basin (Newfoundland, Canada) (Osprey H-84 well; Holser et al 1988). Formations are as follows: 1, unnamed sedimentary formation; 2, Clubhouse Crossroads Basalt; 3, Balls Bluff, Tibbstown, and Catharpin Creek formations; 4, Mt. Zion Church Basalt; 5, Midland Formation; 6, Hickory Grove; 7, Turkey Run Formation; 8, Sander Basalt; 9, Waterfall Formation; 10, Gettysburg Formation; 11, Aspers Basalt; 12, unnamed sedimentary unit; 13, Passaic Formation; 14, Orange Mountain Basalt; 15, Feltville Formation; 16, Preakness Basalt; 17, Towaco Formation; 18, Hook Mountain Basalt; 19, Boonton Formation; 20, South Britain Formation; 21, lower basalt; 22, lower shale and conglomerate; 23, upper basalt; 24, upper shale; 25, New Haven Formation; 26, Talcott Basalt; 27, Shuttle Meadow Formation; 28, Holyoke Basalt; 29, East Berlin Formation; 30, Hampden Basalt; 31, Portland Formation; 32, Sugarloaf Formation; 33, "Fall River beds" of Olsen et al (1992); 34, Deerfield Basalt; 35, Turners Falls and Mt. Toby formations; 36, Bigoudine Formation; 37, Argana Basalt; 38, unnamed sedimentary unit; 39, Blomidon Formation; 40, North Mountain Basalt; 41, Scots Bay and McCoy Brook Formations; 42, Argo Formation and unnamed shale unit; 43, unnamed basalt; 44, Eurydice and Osprey Formation; 45, unnamed basalt; 46, Argo Formation; 47, Argo Formation. Chemical data from Tollo & Gottfried (1992), Puffer (1992), Pe-Piper et al (1992), and Dostal & Greenough (1992).

known for over 100 years (see Lorenz 1988, McDonald 1996). Three lava flow formations (Emerson 1892, Davis 1898, Gray 1982, Philpotts & Reichenbach 1985, Philpotts & Martello 1986, Philpotts 1992, Puffer 1992), two interbedded sedimentary formations, and one thick overlying sedimentary formation make up the section (Figures 3, 18). The lower sedimentary units (Shuttle Meadow and East Berlin formations) are largely lacustrine to marginal lacustrine in origin, and the upper one (Portland Formation) grades gradually upward into wholly fluvial units (Lehmann 1959). Similar though less-complete sequences are seen in the youngest portions of the Deerfield, Pomperaug, Newark, Gettysburg, and Culpeper basins.

The transition from TS III to TS IV in the Hartford basin is marked by a change of provenance of the basin fill, as well as a switch from entirely fluvial to predominately lacustrine deposition with interbedded lava flows, probably due to an increase in extension rate with a concomitant increase in the basin asymmetry (McDonald & LeTourneau 1988). Although the uppermost New Haven Formation might also belong within TS IV, the unambiguously lowest unit in the sequence in the Hartford basin is the Talcott Basalt (Lehmann 1959, Sanders 1962, Sanders 1970), which belongs to the high-titanium quartz-normative (HTQ) group of Puffer (1992). The Talcott Basalt is absent from the northern third of the Hartford basin and may have been cut out by an unconformity (Cornet 1977), or it may never have been deposited there. A feeder system to this formation has been described by Philpotts & Martello (1986) and Philpotts & Asher (1992). Red, gray, and black clastic rocks and limestones of the Shuttle Meadow Formation (Krynine 1950, Lehmann 1959, Hubert et al 1978, 1992, Smoot 1991a, Huber & Gierlowski-Kordesch 1994, Gierlowski-Kordesch 1995) overlie the Talcott Formation. The carbonate units are deep-water parts of Van Houten cycles that mark out a cyclical pattern in the lower half of the formation (Olsen et al 1996b). They are usually very rich in well-preserved fossil fish (Cornet et al 1973, Olsen et al 1982).

The Holyoke Basalt (Lehmann 1959), which overlies the Shuttle Meadow Formation, is a high-iron quartz-normative and incompatible element–depleted tholeiite (HFQ/IED basalt of Puffer 1992) comprising the thickest and most widespread extrusive unit in the Hartford basin. Although the lower third of the next higher formation, the East Berlin Formation, is poorly known, the cyclical upper two thirds has been well described (Lehmann 1959, Hubert et al 1976, 1978, 1982, Demicco & Kordesch 1986, Meriney 1988, Olsen et al 1989, Gierlowski-Kordesch 1991, Gierlowski-Kordesch & Rust 1994, Olsen et al 1996b). In total, the East Berlin Formation consists of a very well-developed sequence of Van Houten cycles comprising three short modulating cycles (Figure 19). In the upper half of the formation there are six black shale–bearing Van Houten cycles that have been traced more or less basin-wide

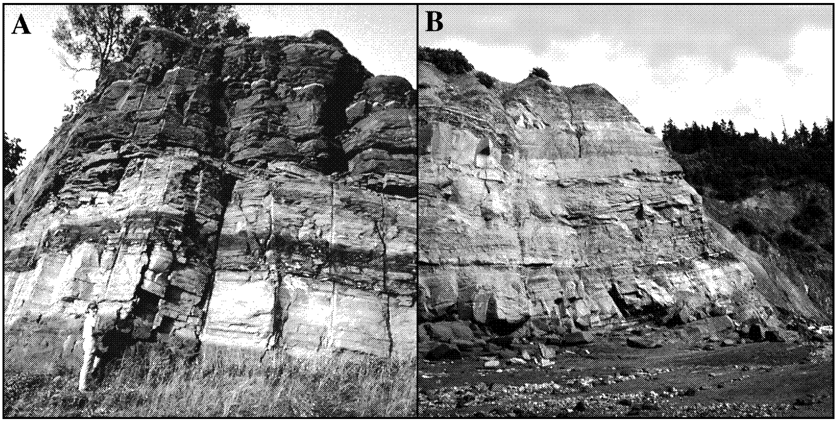


*Figure 19* Examples of TS IV in Hartford (A) and Fundy (B) basins. (A) Single Van Houten cycle in Middle East Berlin Formation, East Berlin, Connecticut, United States (see Figure 20). Base of cycle is at woman's (Annika Johansson) head, top is at the top of the outcrop. The two light-colored beds are predominately shallow-water gray sandstone and mudstone; the dark band between them is a deep-water black microlaminated carbonate and mudstone, and the beds above and below the light-colored sandstones and mudstones are playa and sheet delta red sandstones and mudstones. (B) Fluvial-lacustrine red and brown sandstones and mudstones of the McCoy Brook Formation (*foreground*) with dark-colored breccia and flows of the underlying North Mountain Basalt (here separated by a fault from the overlying formation). Outcrop is at Five Islands Provincial Park, Nova Scotia, Canada, looking east.

(Figure 20) (Hubert et al 1976, Olsen 1985, Olsen et al 1989). There are some beds rich in evaporite pseudomorphs (Demicco & Kordesch 1986, Gierlowski-Kordesch & Rust 1994), and Suchecki et al (1988) report vertical variations in oxygen and carbon isotopic ratios through Van Houten cycles in the East Berlin Formation are consistent with the lake contraction due to evaporation.

The thinnest of the tholeiitic lava flow sequences, the Hampden Basalt (Lehmann 1959) overlies the East Berlin Formation. It belongs to the high-iron, high-titanium quartz-normative (HFTQ) group of Tollo & Gottfried (1992), which is the same as the high-iron, quartz-normative incompatible element–enriched (HFQ/IEE) group of Puffer (1992). Its flow characteristics have been described by Chapman (1965) and Gray (1982), and its feeder system has been described by Foose et al (1968) and Philpotts & Martello (1986). In the northern part of the Hartford basin, the Hampden Basalt is locally replaced by the Granby Tuff (Emerson 1898, April & Keller 1992), its volcanoclastic equivalent.

The Portland Formation comprises the youngest formation in the Hartford basin and the thickest portion of TS IV preserved anywhere ($\sim$2 km). The lower Portland Formation consists of a cyclical, largely lacustrine sequence

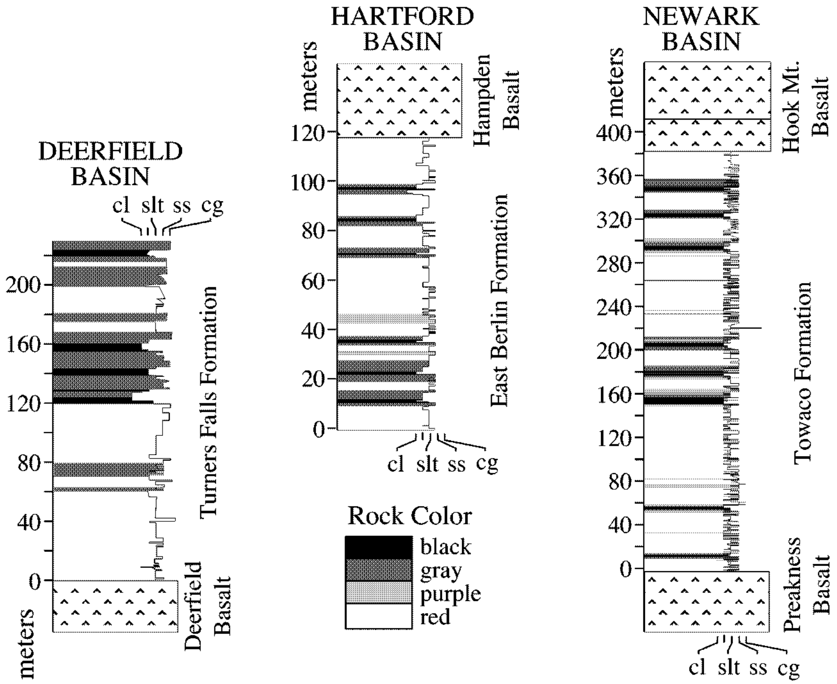


*Figure 20* Comparison of the cyclostratigraphy of three correlative formations in three different basins. The Newark basin section is based on the ACE cores (Olsen et al 1996b), and the Hartford (Olsen et al 1989) and Deerfield basin (Olsen et al 1992) sections are based on single outcrops. Van Houten cycle in Figure 19A is centered at 40 m in the East Berlin section.

very similar in overall aspect to the East Berlin Formation (Emerson 1898, Krynine 1950, Lehmann 1959, Sanders 1968, Gilchrist 1978, LeTourneau 1985, Olsen et al 1989, Smoot 1991a, Pienkowski & Steinen 1995). Beds rich in pseudomorphs of evaporite crystals are locally common (Parnell 1983), although no bedded evaporites have been found. Marginal lacustrine facies have been described by LeTourneau (1985), LeTourneau & Smoot (1985), and McDonald & LeTourneau (1988). According to Smoot (1991a), the cyclical lacustrine sequence passes up into sequences with shallow water lacustrine sequences (McDonald & LeTourneau 1989) and meandering stream deposits, which pass upward into red completely fluvial sequences described by McInerney (1993).

*Other basins* The stratigraphy of TS IV in the Deerfield basin is very similar to that in the northern part of the Hartford basin (Olsen et al 1992) (Figures 18, 20).

There is only one extrusive tholeiitic formation, the Deerfield Basalt (Emerson 1891), and, like the Holyoke Basalt, it is of the HFQ/IED type of Puffer (1992) (Tollo & Gottfried 1992). Underlying the Deerfield Basalt are the "Fall River beds" that are lithologically very similar to the Shuttle Meadow Formation of the Hartford basin (Stevens & Hubert 1980, Olsen et al 1989, 1992), at the base of which may lie an unconformity with TS III (Sugarloaf Formation) (JP Smoot, personal communication). Overlying the basalt is a cyclically lacustrine sequence (the Turners Falls and Mt. Toby formations; Willard 1951, Emerson 1891, Robinson & Luttrell 1985), the basal part of which has a cyclostratigraphy virtually identical to the East Berlin Formation of the Hartford basin and the Towaco Formation of the Newark basin (Meriney 1988, Olsen et al 1989, 1992) (Figure 20). The rest of the Turners Fall and Mt. Toby formations is in its general lithological character and cyclical character very much like the lower Portland of the Hartford basin (Handy 1976, Olsen et al 1989, 1992). An unconformity that supposedly lies within the Turners Falls Formation, in the position of the Hampden Basalt (Cornet 1977, Robinson & Lutrell 1985), apparently does not exist (Olsen et al 1989, 1992).

The TS IV in the tiny Pomperaug basin of Connecticut is similar to that in the Hartford basin, albeit much thinner and lacking the HFTQ flows and overlying units. (Sanders 1963, Lorenz 1988, Huber & McDonald 1992, Hurtubise & Puffer 1983). Offshore of Massachusetts, TS IV has been encountered in the Nantucket basin (Hutchinson & Klitgord 1988) where tilted basalt flows have been described by Folger et al (1978).

The basic stratigraphy of TS IV in the Newark basin was described by Darton (1890) and Kümmel (1897) and later refined by Olsen (1980a,b, 1995) and Olsen et al (1996a,b). The overall stratigraphy is remarkably like that of the Hartford basin in its cyclostratigraphy (Olsen et al 1986, Olsen et al 1996b), overall lithological sequence (Olsen et al 1989), and lava flow chemistry (Puffer & Lechler 1980, Puffer 1992, Tollo & Gottfried 1992) (Figures 18, 20). There is, however, an additional type of basalt, the low-titanium quartz-normative type (LTQ of Puffer 1992) in the middle basalt formation (Preakness Basalt). Volcanic structures have been described by Darton (1890), Manspeizer (1980), Olsen (1980b), Faust (1975), and Puffer & Student (1992). The Palisades sill was a feeder to the basalt flows in the Newark basin (Ratcliffe 1988), and the $^{40}Ar/^{39}Ar$ and U-Pb dates from the sill of about 201 Ma (Sutter 1988, Dunning & Hodych 1990) are the best available for the Newark basin flows (but see Seidemann 1989, 1991).

Virtually all of TS IV is represented in the ACE cores (Fedosh & Smoot 1988, Olsen et al 1996b), and the lower two formations were recovered by the NBCP cores (Olsen et al 1996a). The cyclostratigraphy of this sequence is therefore known better than in any other basin (Olsen et al 1996b). Sedimentation rates increased dramatically from the underlying Passaic Formation into the extrusive zone and then declined afterward (Figure 6). The increase in accumulation rate is mirrored by a dramatic increase in the frequency of well-developed black shales and carbonates in Van Houten cycles. The increase in the degree of development of lake high-stand sequences occurs about 500 ky prior to the base of the first basalt and includes the Triassic-Jurassic boundary, which suggests that the base of TS IV lay within the uppermost Passaic Formation (Figures 6, 16). Interestingly, the pattern of carbonate-rich Van Houten cycles in the lowest Jurassic formation (Feltville) passing up into more siliclastic Van Houten cycles by the middle of the next Jurassic formation (Towaco) is very much like that seen in the Hartford basin (Olsen 1980c, Olsen et al 1992). Abundant and diverse fish from the deeper water portions of Van Houten cycles from this tectonostratigraphic sequence have provided evidence for very rapid speciation in geologically ephemeral lakes (McCune et al 1984, McCune 1990, 1996, Olsen & McCune 1991).

TS IV is represented in the Gettysburg basin by only two small areas of Jurassic-age lava flows of HTQ type (Stose & Bascom 1929, Smith et al 1975) and associated lacustrine and alluvial rocks (Cornet 1977).

The Culpeper basin is the southernmost area in the Newark Supergroup known to have extrusive flows interbedded with extensive sedimentary formations, and it is thus the most southern to have a typically developed TS IV. The stratigraphy has been described by Cornet (1977), Lee (1978, 1979), Lee & Froelich (1989), and Lindholm (1979). As in the Hartford, Pomperaug, and Newark basins, there are three tholeiitic basalt formations that exhibit an upward sequence of changes in chemistry from HTQ in the lowest formation (Mt. Zion Church Basalt) to HFQ/IED and then LTQ and HFQ/IED in the higher flow formations (Hickory Grove and Sander Basalts) (Tollo 1988, Puffer 1992). However, there does not seem to be a flow with the HFTQ type chemistry. Again, like the Hartford, Pomperaug, and Newark basins, the Van Houten cycles of the lowest Jurassic-age sedimentary formation (Midland Formation) are carbonate-rich (Gore 1994, Roberts 1989, Roberts & Gore 1988, Gore 1988c), whereas the upper ones (Turkey Run, Waterfall) are more siliclastic (Hentz 1985, Olsen et al 1989, Gore 1988c). Van Houten cycles in the Waterfall Formation resemble those of the Towaco Formation near the border fault system in the Newark basin (see Olsen et al 1996b), although they are individually on the order of 150 m thick (Hentz 1985, Gore 1988c), compared to 25 m in the Towaco Formation.

Apparently untilted basalt flows lie on sedimentary sequences that are probably TS III and prerift basement beneath the coastal plain and on the continental shelf of South Carolina, Georgia, and Florida (Gohn & Higgins 1978, McBride et al 1989). The Clubhouse Crossroads Basalt of the South Georgia rift (Figure 1) was drilled in South Carolina and consists of multiple flows with very little intercalated sediment. Geochemically, the basalts are HFQ/IED and olivine normative (Gottfried et al 1983, Puffer 1992) (the latter may be a result of alteration; PC Ragland, personal communication); the south Florida basalts are HFQ/IED type (Puffer 1992). These flows were most likely extruded contemporaneously with those in the more northeastern United States (Olsen et al 1996b) but apparently after the southern rift basins had ceased to subside.

The uppermost Bigoudine Formation with the Triassic-Jurassic boundary, the Argana Basalt, and the overlying sedimentary sequence in the Argana basin belong to TS IV in a sequence closely comparable to that in the Fundy basin. The Argana and related early Mesozoic tholeiitic basalts of Morocco (Manspeizer et al 1978) are similar in composition to the HFQ basalts (Bertrand et al 1982, Fiechtner et al 1992). There are some thin intercalated sediments between individual flows. In at least the Argana basin, the basalts are overlain by an unnamed thin unit of cyclical red, gray, and black mudstones and carbonates (JP Smoot, personal communication), which is separated from the overlying Early Jurassic marginal marine deposits by an unconformity (Manspeizer 1988; PE Olsen, personal observation). To the west and north in Morocco, the basalts tend to be overlain (as well as underlain) by evaporites in relationships more comparable to those seen on the Scotian Shelf and Grand Banks (Pe-Piper et al 1992).

The TS IV is very well represented in the Fundy basin, consisting of the uppermost Blomidon Formation, the North Mountain Basalt (Powers 1916), and the overlying Scots Bay (Powers 1916) and McCoy Brook formations (Donohoe & Wallace 1978) (Figure 19). The uppermost Blomidon Formation contains the Triassic-Jurassic boundary and the only known dark shale beds within the formation. The overlying North Mountain Basalt is an HTQ tholeiite that has been cored almost in its entirety (Getty Mines cores; Cameron & Jong 1985) and drilled through by two industrial exploratory wells (Pe-Piper et al 1992). The multiple flows have been described in distribution, morphology, and geochemistry by Powers (1916), Klein (1962), Liew (1976), Colwell (1980), Olsen (1981), Papezik et al (1988), Olsen et al (1989), Olsen & Schlische (1990), Pe-Piper et al (1992), Dostal & Greenough (1992), Greenough & Dostal (1992a,b), Greenough (1995), and Schlische & Ackermann (1995). A U-Pb data of 202 Ma has been obtained from the North Mountain Basalt (Hodych & Dunning 1992), which is in agreement with its geochemical correlatives in the Newark basin.

The overlying formation is locally a thin sequence of gray and white limestone and chert that is called the Scots Bay Formation (Powers 1916), which, according to De Wet & Hubert (1989) and Suchecki et al (1988), was deposited in part by hot springs arising from the basalt. The lateral equivalent of the Scots Bay Formation is the mostly red McCoy Brook Formation (Donohoe & Wallace

1978). Throughout most of its exposed extent, the formation is shallow-water lacustrine, marginal lacustrine, and fluvial. There are minor gypsum-rich intervals and limestones, and adjacent to the strike-slip margin of the basin there are significant eolian dune sands and basalt talus slope deposits (Olsen et al 1989, Olsen & Schlische 1990, Tanner & Hubert 1992). The latter have produced a very rich vertebrate assemblage (Olsen et al 1988). Based on the relative rarity of evaporites and sand patch mudstone, and the abundance of bioturbation and fluvial-lacustrine deposits, the McCoy Brook and Scots Bay formations were deposited under generally more humid conditions than the bulk of the underlying Blomidon Formation, which the McCoy Brook Formation superficially resembles.

The TS IV is represented by tholeiitic basalts and overlying synrift sedimentary strata that have been encountered in several wells on the Scotian Shelf and on the Grand Banks (Pe-Piper et al 1992) (Figure 1). In the Glooscap C-83 well on the Scotian shelf, basalt of HTQ and olivine-normative types (Pe-Piper et al 1992) is unconformably overlain by Pliensbachian and younger age marine units. The Spoonbill C-30 and the Cormorant N-83 on the Grand Banks have HTQ-type basalts overlain by evaporites (Pe-Piper et al 1992). In the majority of basins on the Scotian Shelf and Grand Banks, there are no basalts and the boundary between TS III and TS IV lies unrecognized, and perhaps unrecognizable, either within Argo Formation evaporites or Eurydice Formation mudstones, or between them.

EUROPE THROUGH GREENLAND Triassic and Early Jurassic sequences in Iberia, Europe, Great Britain, Norway, Greenland, and Svalbard have much in common with the CAM basin sequences described here, although assigning causal relationships is still very speculative. In particular the "Bunter" sequences of Europe and further north, which were deposited during a period of significant rifting (Ziegler 1988), may be related to TS I. The TS II in the CAM basins may be related to contemporary deposits in Europe (e.g. Schilfsandstein). The younger Keuper, especially the Knollenmergel, Mercia Mudstone, and similar deposits in Europe, Great Britain, Norway, and Greenland, appear more sag-like in their geometry and probably postdate early Mesozoic rifting. Demonstration that these sequences at least correlate in time to TS III is provided by Kent & Clemmenson's (1996) detailed magnetostratigraphic correlation of the Fleming Fjord Formation of East Greenland with the middle and upper Passaic Formation of the Newark basin. Such correlations will allow much more refined analysis of intercontinental scale relationships. Finally, there is probably a causal link between the marine transgressions of the Early Jurassic and the accelerated subsidence and concomitant accumulation seen in TS IV (e.g. Hallam 1992).

## LINKAGES BETWEEN BASINS

After a period in which the map geometry of the North American CAM basins was determined, it became apparent that each basin is comprised of one or more half graben, in large part defined by a major fault system, usually on one side of the basin, toward which the basin strata generally dip (see summary in Russell 1892). Russell (1878) noted that the basins seemed arranged in paired lines of basins of opposing symmetry with similar stratigraphies, and suggested that all of the basin areas represent the remnants of one formerly continuous full graben that was anticlinally folded after deposition and deeply eroded. This concept, called the broad terrain hypothesis, was controversial from the start and has been invoked, periodically, for over 100 years, most recently by McHone (1996). The opposing theory, that the basins formed more or less as they are presently disposed, has been called the isolated basin hypothesis (see review by Lorenz 1988).

Whereas it is now clear (as discussed above) that half graben were integral to CAM basin geometry throughout their depositional history, the basins and basin complexes preserved in the subsurface are less deeply eroded than those that are exposed (Figure 7) and show more complex relationships than would be suggested by either end-member hypothesis. In particular, the South Georgia rift (McBride 1991) and the Taylorsville basin (LeTourneau 1996) are half graben complexes connected by thinner sequences of the younger strata (Figure 7). If they were more deeply eroded, these basins would appear, like those already exposed, as isolated half graben. Paleontological data that argued for isolation of the basins of the Newark Supergroup (Olsen et al 1982) were overturned by new data (Olsen 1983). It therefore appears likely that the Newark Supergroup basins consisted of half graben, at least some linked late in their history by a veneer of young (but still synrift) strata, much like the modern Basin and Range half graben (McHone 1996a,b) but unlike the more isolated basins of the East African rift system (Rosendahl et al 1986). The same observations and concepts apply to the numerous basins and basin complexes of Morocco (Van Houten 1977, Manspeizer et al 1978, Manspeizer 1982).

Lambiase (1990) proposed a model of rift basin evolution in which the same facies pattern seen in most of the tectonostratigraphic sequences in the CAM basins is explained by relay relationships of filling between hydrographically linked basins. Although the model provides no quantitative predictions for the filling sequences in individual basins, it does potentially explain the facies relationships between some adjacent basins. In particular, it could provide a framework for relating the Hartford basin Triassic-age fluvial sequence to the adjacent and contemporaneous lacustrine units in the Newark basin. A reasonable hypothesis would be that the Hartford and Deerfield basins were upstream from a major river entering the Newark basin and was trapping most of the clastics of the drainage basin against the sill between the Newark and Hartford basins (i.e. accommodation zone). The same relationship may hold between the Culpeper, Gettysburg, and Newark basins during the deposition of the largely fluvial Manassas Formation, the middle and upper New Oxford Formation of the Gettysburg basin, and the contemporaneous lacustrine Lockatong Formation of the Newark basin. Likewise, we anticipate linkages between the subbasins of the Deer River basin and between the Richmond and Taylorsville basins. A quantitative prediction of these hypotheses is that the timing of the linkages between the drainages should coincide with an increase in accumulation rate in the downstream basin.

Basin histories might also be linked through relay relationships in extension. An example might be the Taylorsville and Culpeper basins (Figure 1). The two basins lie parallel to each other with their midpoints lining up along the inferred direction of extension. Preliminary data from the Taylorsville basin suggests that lacustrine deposition began in the Culpeper basin while deposition in the Taylorsville basin was winding down (Malinconico 1995). Because they were presumably in the same regional stress field, extension may have been abandoned in the Taylorsville basin and picked up in the Culpeper basin, leaving the regional amount of extension the same but distributing the stress in separate lithotectonic terranes as extension progressed. A similar process could happen along strike, making the relationship between regional stress changes and basin-scale extension rate changes quite difficult to decipher.

## DRIFT THROUGH DIFFERENT CLIMATE BELTS

Although the tectonostratigraphic sequences segment the CAM basin sections into sequences that follow relatively simple trajectories in facies that seem to be mostly under the control of tectonic events, there is still a major component of vertical changes that is best explained as a result of the drift of Pangea through more or less stationary zonal climate belts. This relationship, first outlined by Robinson (1973) for the early Mesozoic, has only recently become clear because of new, strongly corroborated pole positions for North America (Kent et al 1995, Kent & Olsen 1996). Most of the apparent changes in overall climate recorded in these basins can thus be attributed to the drift of Pangea.

The changing position of depositional basins fixed on Pangea can be usefully viewed in a space-time diagram (Figure 15). Assuming minimal changes through time in global climate, different basins can be examined along lines of equal paleolatitude and thus, as nearly as possible, equal climates. Or, a single transect through climate belts can be looked at within a single time window. A narrow (2°) swath is visible in which coals are preserved, north of which is a 7° region in which cyclical black shales were deposited. Farther north is a zone in which arid facies dominate.

This matrix of geographically fixed basins on the North American and African plates moving through different paleolatitudes extends the latitudinal range of the climate "samples" over 10° during the Late Triassic. Thus, we can look at cyclical lake sediments in the Newark basin responding primarily to the ∼20-ky cycle of climatic precession at 4°N and look at the same age cyclical lake deposits in the Dan River basin formed at the equator. The latter responded to the double precession cycle. If we look again at the Newark basin 10 my later when it was at 8°N, we see cyclical lake beds still formed in response to the ∼20-ky cycle of climatic precession. Looking 2.5°S of the Newark basin to the Culpeper basin, we see cyclical lake deposits responding to virtually the same rhythm. If the Dan River basin preserves a lacustrine sequence of this age (which it may), one would predict the presence of only the ∼20-ky cycle of climatic precession, because by then the basin was at 4°N. At the same time, however, the Blomidon Formation of the Fundy basin was responding to the 40-ky obliquity cycle. Similarly, we might predict that although the earliest Jurassic strata of the Hartford basin responded to the ∼20-ky climatic precession cycle, the younger Jurassic strata in the Hartford basin may well have drifted north into the zone dominated by the 40-ky obliquity cycle. These strata have not yet been examined with this in mind.

Obviously this method highlights the need for careful and rigorous temporal correlations and paleogeographic analysis. In fact, because of the high climatic gradients around the Pangean equator and variable rates of movement of the supercontinent, it is essential that both paleogeography and time be as closely constrained as possible if ancient geological data are to be used for meaningful tests of climate models. Thus far, much ad hoc speculation on differences between theoretical and numerical predictions and geological data has been expended on jumbles of noncontemporaneous bits of climate-sensitive data of variable quality in the analysis of Pangean climate. In fact, although discussion of Pangean climate has most recently been focused on its nonzonal aspects (Manspeizer 1982, Parrish et al 1982, Wilson et al 1994), Late Triassic climate appears quite zonal with the methods outlined here and more like the models of Kutzbach (1994), with rather sharp gradients. In fact, the Late Triassic was a time of comparatively very slow drift. Considering such conditions coupled with the high-resolution stratigraphic tools that are now available, Late-Triassic "hothouse" Pangea would seem uniquely suited for climate testing studies as exemplary of an extreme end-member of Earth's climate and geography states.

## REGIONAL AND GLOBAL CLIMATE EVENTS

Although much of the variability in climate-sensitive rocks is compatible within this framework of slow northerly drift, some of the variability is not. Stated more specifically, just before the Triassic-Jurassic boundary, the frequency of more humid facies increases dramatically across all latitudes regardless of tectonic setting along this more or less meridian tract. Yet, North America should be drifting farther into the arid belt during this time. The direction of apparent climate change is thus at variance with the predictions of drift. In addition, this very rapid change in apparent climate came at a time of very slow drift (Kent et al 1995). There is no sign of an abrupt increase in wetness in the overall drying trend seen in the early Mesozoic deposits of the western United States, and thus the change does not seem to be global. One simple explanation is that the climate became more humid regionally, although within the CAM basins some of the trend could be due to increased extension marking the base of TS IV. It is interesting to speculate whether this humidification could have been associated with the Rhaetian–Early Jurassic transgression, marked in the northern part of the CAM basins by the appearance of very widespread marine-derived evaporites. On the other hand, the Triassic-Jurassic boundary itself was one of the largest mass extinctions of all time (Hallam 1992, Olsen et al 1988, Sepkoski 1993). The CAM basins are one of the principle and highest resolution recorders of the event (Olsen et al 1988, Olsen et al 1990, Silvestri & Szajna 1993, Fowell & Olsen 1995, Fowell et al 1994). Explanations for the mass extinction run from the nearly contemporaneous flood volcanism (Courtillot 1994), to marine transgressive-regressive couplets (Hallam 1992), to asteroid or comet impact (Dietz 1986, Olsen et al 1988, Bice et al 1992). Although the ∼100-km Manicouagan impact structure of Quebec (Olsen et al 1988) had been identified as a candidate impact structure for the boundary, recent studies indicate that the impact is too old (Hodych & Dunning 1992). Nonetheless, at present, circumstantial evidence favors an impact origin for the mass extinction because both a “fern spike” (Figure 16) and shocked quartz (Fowell et al 1994, Bice et al 1992) have been found at the boundary. The climate change evident near the boundary in the CAM basins, as well as the flood basalt event that just postdates it, could be part of the causal chain or could be coincidental.

## CONCLUSIONS

The CAM basins are the result of the incipient rifting of Pangea and preserve a record spanning more than 30 million years of the Late Triassic and Early Jurassic. They provide a natural laboratory for unraveling the interplay between climatic and tectonic records in continental rifts. An especially valuable aspect of the rift fill is the pervasive sedimentary cyclicity produced by Milankovitch-type forcing of climate recorded in high–accumulation rate lacustrine sequences. These strata allow calibration of many processes within the basins, as well as the production of a Triassic–Early Jurassic time scale.

Nonetheless, there are many critical unanswered questions, a few of which are: 1. What is the relationship between flood basalt extrusion, climate change, and extinctions? 2. What is the relationship between the geographic pattern of volcanism, intrusion, and cessation of basin subsidence? 3. What controls partitioning of the section into tectonostratigraphic sequences? 4. How do we calibrate the altitude of the depositional surface of the basins (e.g. Manspeizer 1982)? 5. What is the theoretical underpinning of the geographic variations in Milankovitch climate forcing around the equator? 6. What are the main physical processes that will allow us to quantitatively predict basin subsidence patterns? 7. How are the climate cycles that produced the dramatic lacustrine cyclicity reflected in contemporaneous fluvial sequences? 8. How can we tie the detailed time scale developed in the Newark basin (Kent et al 1995) to the much less stratigraphically complete marine records and to the ocean-floor magnetic anomaly time scale that ends in the Middle Jurassic? 9. Was the Triassic-Jurassic boundary really a consequence of a bolide impact, and did the Manicouagan impact really produce no biological effects?

The Triassic and earliest Jurassic were times of global hothouse conditions, during which the continents were united in the supercontinent of Pangea. This period was a critical time in Earth history and one of the Earth's greatest natural experiments. New techniques that allow for high-resolution stratigraphy and paleogeography as well as new understanding of the role of celestial mechanics in cyclostratigraphy are beginning to allow for wholly new ways of treating these ancient Pangean sequences. No a priori reasons prevent us from now asking questions of these ancient sequences that require the same level of resolution that we have so far reserved for the Neogene.

## ACKNOWLEDGMENTS

I thank Sarah Fowell, Dennis Kent, Peter LeTourneau, Mary Ann Malinconico, David Reynolds, Roy Schlische, and Martha Withjack for many pleasant hours of discussion on rift basin topics that have substantially influenced this paper—although I take full responsibility for any errors of fact or omission. I am grateful for reviews of the manuscript by Annika K. Johansson, Peter M. LeTourneau, and Mary Ann Malinconico, who substantially improved it. This work was supported by National Science Foundation grant ATM 93-17227. This is LDEO contribution no. 5576.

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