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QUANTITATIVE FILLING MODEL FOR CONTINENTAL EXTENSIONAL BASINS WITH APPLICATIONS TO EARLY MESOZOIC RIFTS OF EASTERN NORTH AMERICA[1]

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ABSTRACT

In many half-graben, strata progressively onlap the hanging wall block of the basins, indicating that both the basins and their depositional surface areas were growing in size through time. Based on these constraints, we have constructed a quantitative model for the stratigraphic evolution of extensional basins with the simplifying assumptions of constant volume input of sediments and water per unit time, as well as a uniform subsidence rate and a fixed outlet level. The model predicts (1) a transition from fluvial to lacustrine deposition, (2) systematically decreasing accumulation rates in lacustrine strata, and (3) a rapid increase in lake depth after the onset of lacustrine deposition, followed by a systematic decrease. When parameterized for the early Mesozoic basins of eastern North America, the model's predictions match trends observed in late Triassic-age rocks. Significant deviations from the model's predictions occur in Early Jurassic-age strata, in which markedly higher accumulation rates and greater lake depths point to an increased extension rate that led to increased asymmetry in these half-graben. The model makes it possible to extract from the sedimentary record those events in the history of an extensional basin that are due solely to the filling of a basin growing in size through time and those that are due to changes in tectonics, climate, or sediment and water budgets.

INTRODUCTION

A fundamental problem in the study of extensional basins is the quantitative determination of the first-order controls on basin stratigraphy. This problem is especially acute for non-marine basins, which lack a sea-level datum. Traditionally, each major change in depositional environment and sedimentation has been interpreted as a result of some "tectonic" event (such as increased boundary fault movement) or major climatic change, yet predictive, quantitative models for the stratigraphic development of extensional basins are notably lacking. This stands in contrast to the study of post-rift deposits, in which the fundamental aspects of the stratigraphy are explained in terms of thermally driven subsidence, sediment loading, sediment compaction, and eustatic sea level change (e.g., McKenzie 1978; Steckler and Watts 1978; Bond and Kominz 1984; Steckler et al. 1988).

Quantitative stratigraphic modeling of extensional basins has proven difficult because of the lack of appreciation of the critical effects of basin geometry on the stratigraphic record. In this paper, we develop a simple model of extensional basin filling that appears to explain many of the elements of the stratigraphic record of early Mesozoic rift basins of eastern North America. The model is based on the filling of a basin growing in size through time. Even under conditions of uniform subsidence and constant volume of sediment and water input per unit time, the model

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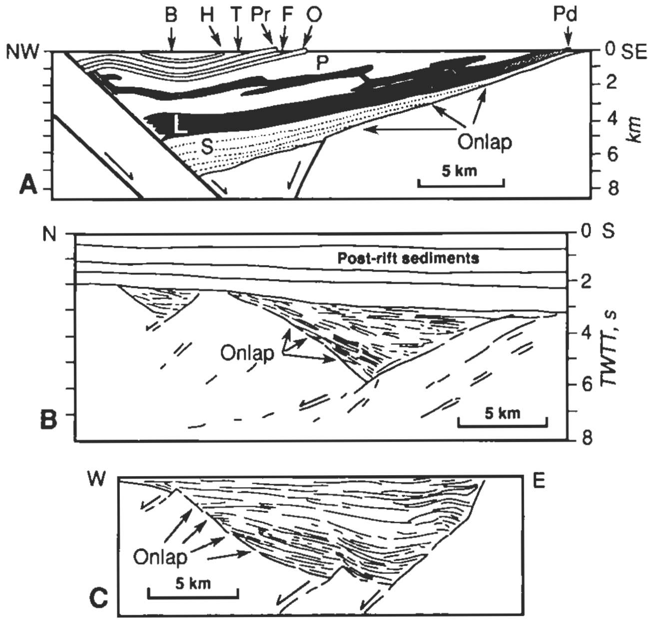

FIG. 1.—Cross section and seismic lines of typical half-graben, showing the wedge-shaped geometry and the progressive onlap of younger strata onto the hanging wall basement block on the hinged side of the (*A*) Newark (see figs. 5 and 7 for location), (*B*) Atlantis (see fig. 5 for location), and (*C*) Railroad Valley basins. Abbreviations for stratigraphic units in the Newark basin are: *B*, Boonton Formation; *F*, Feltville Formation; *H*, Hook Mountain Basalt; *L*, Lockatong Formation; *O*, Orange Mountain Basalt; *P*, Passaic Formation; *Pd*, Palisades diabase; *Pr*, Preakness Basalt; *S*, Stockton Formation; and *T*, Towaco Formation. *TWTT* is two-way travel time. Seismic profiles were traced from figures presented by Hutchinson et al. (1986) and Vreeland and Berrong (1979).

predicts transitions in major depositional environments and trends in accumulation rates and lake depth. Furthermore, the model's predictions can be "subtracted" from the actual stratigraphic record, yielding the tectonic and climatic signal and allowing for the parameterization of sediment and water budgets.

EXTENSIONAL BASIN GEOMETRY AND STRATIGRAPHIC ARCHITECTURE

In many extensional basins, younger strata progressively onlap the "basement" rocks of the hanging wall block (fig. 1). Charles Lyell (1847) first noted this onlap geometry in the Blackheath region of the hinge area of the Triassic Richmond basin of Virginia. Outcrop studies and coal mine excavations indicate that the lowermost stratigraphic unit—the Lower Barren Beds—is absent from this region and that the stratigraphically higher Productive Coal Measures rest directly on basement. Further toward the border fault of the half-graben, seismic reflection profiles and drill hole data record the presence and thickening of the Lower Barren Beds (B. Cornet pers. comm. 1989). At Tennycape, Nova Scotia, strata of the Triassic Wolfville Formation clearly onlap the Carboniferous "basement" rocks in the hinge of the Fundy half-graben. Furthermore, the beds of the Wolfville Formation dip at a shallower angle than the unconformity between the Triassic and Carboniferous rocks (Olsen et al. 1989). McLaughlin stated that the basal Stockton Formation onlapped the northeastern, southern, and southwestern margins of the Newark basin (McLaughlin 1945; McLaughlin and Willard 1949; Willard et al. 1959; see fig. 1*A*). Seismic reflection profiles reveal this onlap geometry in (1) the Cenozoic basins of the Basin and Range—the Dixie Valley basin, Nevada (Anderson et al. 1983, their fig. 4), the northern Fallon basin of the Carson Desert, Nevada (Anderson et al. 1983, their fig. 5), the Diamond Valley basin, Nevada (Anderson et al. 1983, their fig. 6), the Railroad Valley basin, Nevada (fig. 1*C*; Vreeland and

Berrong 1979, their fig. 8), and the Great Salt Lake basin, Utah (Smith and Bruhn 1984, their fig. 10); (2) Lake Tanganyika of East Africa (Burgess et al. 1988, their fig. 35-12); (3) the North Viking graben of the North Sea (Badley et al. 1988, their figs. 6 and 8); (4) the Hopedale and Saglek basins of the Labrador margin (Balkwill 1987, his figs. 9 and 11); and (5) the Mesozoic basins of the U.S. Atlantic passive margin—the Long Island, Nantucket, and Atlantis basins (fig. 1*B*; Hutchinson et al. 1986, their figs. 3, 7, 11, and 15).

Anderson et al. (1983) cited this onlap pattern as evidence that the extensional basins had grown in size through time. Leeder and Gawthorpe (1987) related the onlap pattern to the migration of the basin's hinge or fulcrum (reflecting the line basinward of which the hanging wall has experienced subsidence) away from the border fault and have incorporated this concept into their tectono-sedimentary facies models. We agree with these authors' interpretations and extend them one step further: not only were the extensional basins growing in size, but the depositional surface also was growing in area through time as a consequence of the filling of the basin. This conclusion serves as the foundation of the extensional basin filling model.

Continental extensional basins often display a tripartite stratigraphic architecture (Lambiase 1990) which, from bottom up, consists of: (1) a fluvial unit indicating through-going drainage and open-basin conditions; (2) a lacustrine unit with deepest-water facies near its base, gradually shoaling upward, indicating predominantly closed-basin conditions; and (3) a fluvial unit reflecting once again through-going drainage.

These major changes in depositional environments (on a basin-wide scale) have been explained in terms of changes in basin subsidence (Lambiase 1990). The fluvial unit is interpreted to have been deposited during initial, slow basin subsidence, perhaps in a number of linked, small sub-basins. The deep-water lacustrine unit is interpreted to reflect a deepening of the basin resulting from increased subsidence and the coalescence of sub-basins. The upward-shoaling is interpreted to reflect the gradual infilling of the basin, with fluvial sedimentation returning after the basin had filled to the lowest outlet of the basin. Taking a different approach, Olsen and Schlische (1988*a*, 1988*b*) have demonstrated how evolving basin geometry as a consequence of filling under conditions of uniform subsidence and uniform rate of input of sediments and water could produce much the same stratigraphic architecture and in addition make specific predictions about the changes in accumulation rates and times of transitions from one environment to another.

EXTENSIONAL BASIN FILLING MODEL

The analytically simplest model consists of a basin, trapezoidal in cross-section, bounded by planar faults that dip at equal angles, and bounded along-strike by vertical surfaces (fig. 2). In this full-graben model, uniform extension results in uniform subsidence along the boundary faults (see fig. 3), and hence the depth (D) of the basin increases linearly. To maintain simplicity in the model, the volumetric sedimentation rate (V_s, the volume of sediment added to the basin per unit time) and the available volume of water (V_w) are constant. The outlet of the basin also is assumed to be held fixed with respect to an external reference system.

If the basin subsides slowly enough or V_s is large enough, the basin initially fills completely with sediments, and excess sediment and water leave the basin (fig. 3*A*.1). We characterize this mode of sedimentation as fluvial; the basin is sedimentologically and hydrologically open. During this phase, accumulation rate (thickness of sediment deposited per unit time) is constant and equal to the subsidence rate: with an excess of sediments available, the thickness of sediment deposited is governed solely by the space made available through subsidence. As the graben continues to grow, the same volume of sediment added per unit time is spread out over a larger and larger depositional surface area. Ultimately, a point is reached when the sediments exactly fill the basin (fig. 3*A*.2). Thereafter, the sediments no longer can completely fill the growing basin. The basin is sedimentologically closed (all sediments entering the basin cannot leave the basin), and a lake can occupy that portion of the basin between the sediment surface and the basin's outlet (fig. 3*A*.3). After the onset of lacustrine deposition, accumulation rates decrease systematically, driven by the increasing size of

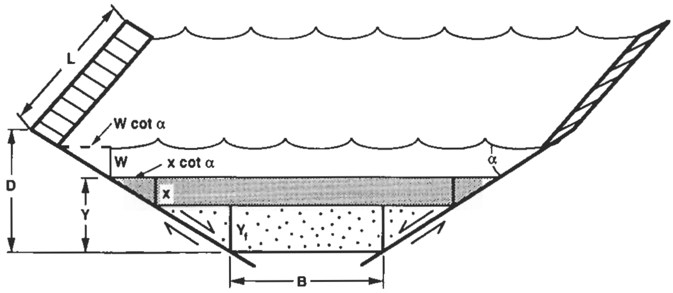

FIG. 2.—Full-graben model, showing meaning of symbols used in the equations in the text.

the depositional surface and constant rate of input of sediments (fig. 3*A*.3–6).

Assuming an unlimited supply of water, lake depth (W) would be equal to the difference between the height of the basin's outlet above the floor of the graben (defined as the basement-sediment contact) and the cumulative sediment thickness (fig. 3*A*.3). However, the supply of water is not infinite, and a point is reached when the graben has grown to such a size that the finite volume of water just occupies the volume between the basin's outlet and the depositional surface (fig. 3*A*.4). Thereafter, the basin is hydrologically closed, and lake depth decreases as a function of the growing size of the basin and nonlinear losses to evaporation from a lake whose surface area is growing through time (fig. 3*A*.5–6).

If the subsidence rate is fast enough or V_s is small enough, the basin never fills completely with sediments, and lacustrine deposition can occur from the outset (fig. 3*B*.1). As the graben grows in size through time, the rate of increase of cumulative sediment thickness and the accumulation rate both decrease. Lake depth increases quickly until the finite volume of water just fills the graben (fig. 3*B*.2) and thereafter decreases (fig. 3*B*.3).

After the onset of lacustrine deposition, accumulation rate should decrease indefinitely. However, graben do not continue to subside indefinitely. If the subsidence rate slows sufficiently or stops and V_s remains constant, the accumulation rate and lake depth would continue to decrease until the depositional surface could reach the basin's outlet, thereby heralding the return of fluvial sedimentation at a rate equal to the subsidence rate.

Quantitatively, under conditions of fluvial sedimentation, cumulative sediment thickness (Y) and the depth of the graben (D) are given by:

$$Y = D = S_R t \qquad (1A)$$

where S_R is the subsidence rate and t is time. Accumulation rate (S) is the derivative of the cumulative thickness curve:

$$S = \mathrm{d}Y/\mathrm{d}t = S_R \qquad (2A)$$

After the onset of lacustrine sedimentation (sedimentologic closure), cumulative sediment thickness and accumulation rate are given by:

$$Y = Y_f + \frac{\tan\alpha}{2} \times \left[\sqrt{(B + 2Y_f\cot\alpha)^2 + \frac{4V_s\cot\alpha}{L}\,t_l} - (B + 2Y_f\cot\alpha)\right] \qquad (1B)$$

$$S = \frac{\mathrm{d}Y}{\mathrm{d}t_l} = \frac{V_s}{L} \times \left[(B + 2Y_f\cot\alpha)^2 + \frac{4V_s\cot\alpha}{L}\,t_l\right]^{-1/2} \qquad (2B)$$

where Y_f is the thickness of fluvial sediments, B is the width of the base of the graben, L is the length of the graben, α is the dip of the border faults, and t_l is time after the onset of lacustrine deposition. While the basin is hy-

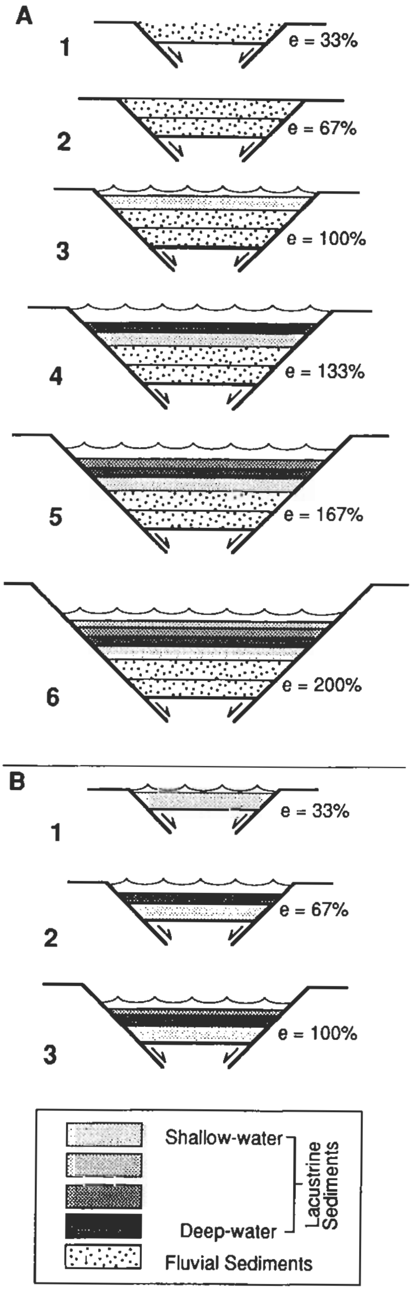

FIG. 3.—Diagrammatic representation of the salient features of the filling of a full-graben under conditions of uniform subsidence and constant volume input of sediment per unit time; e equals percent extension. See text for further explanation.

drologically open, water depth (W) is given by:

$$W = D - Y \tag{3A}$$

Equation (3A) also applies in the case of fluvial sedimentation and correctly predicts zero lake depth since $D = Y$. The equation that describes lake depth after hydrologic closure is:

$$W = \frac{\tan\alpha}{2L} \times \sqrt{(BL + 2YL\cot\alpha)^2 + 4LV_w\cot\alpha} - \frac{B + 2Y\cot\alpha}{2\cot\alpha} \tag{3B}$$

where V_w is the volume of water available.

The derivation of equations (1B) and (3B) are as follows, using the geometry depicted in figure 2. After each increment of subsidence [equation (1A)], the capacity of the basin (the volume of the trapezoidal trough) is calculated. If the capacity is less than the volume of sediment that can be supplied in that increment, the basin is filled with sediments, and the thickness of sediments deposited in that increment is equal to the amount of subsidence for that increment. If the capacity of the basin exceeds the volume of sediment that can be supplied in that increment, the basin cannot completely fill. An integrated approach can then be applied to derive the expression for cumulative sediment thickness [equation (1B)]. From the geometry depicted in figure 2, the volume (V) of sediments is given by

$$V = V_s t_l = (x\cot\alpha)(x)(L) + (B + 2Y_f\cot\alpha)(x)(L \tag{4}$$

where V_s is the volumetric sedimentation rate, t_l is the time after the onset of lacustrine sedimentation, α is the dip of the border faults, L is the length of the basin, B is the initial fault spacing, Y_f is the thickness of fluvial sediments, and x is the next increment of lacustrine sediment thickness. Rearranging terms yields:

$$(\cot \alpha)x^2 + (B + 2Y_f \cot \alpha)x - \frac{V_s t}{L} = 0 \quad (5)$$

The positive root of x is obtained from the quadratic formula. Equation (1B) is then given by

$$Y = Y_f + x \quad (6)$$

After the onset of lacustrine deposition, the capacity of the basin is again calculated knowing Y. If it is less than the available volume of water (V_w), water depth (W) is calculated as the difference between the total depth of the basin and Y. If the basin's capacity exceeds V_w, lake depth is derived as follows, again based on the geometry in figure 2.

$$V_w = (B + 2Y \cot \alpha)(W)(L) + (W \cot \alpha)(W)(L) \quad (7)$$

$$(L \cot \alpha)W^2 + (BL + 2Y \cot \alpha)W - V_w = 0 \quad (8)$$

Solving for the positive root of W yields equation (3B).

In principle, the above analysis should hold for any concave-upward basin in which the depositional surface area grows through time under conditions of uniform subsidence, constant rate of input of sediments and water, and constant position of the outlet fixed to an external reference system, although the form of the equations will differ depending on the evolving geometry of the basin. In a half-graben growing in volume through time (inferred from onlap geometry) under conditions of uniform extension, it therefore seems reasonable to assume that the changes in average accumulation rate and the time of transition from fluvial to lacustrine sedimentation should follow a similar path as for a full-graben as long as the area of the depositional surface continues to grow. Although three-dimensional modeling of the filling of a half-graben has not yet been undertaken, preliminary two-dimensional modeling indicates that the above assumption may be valid. In the two-dimensional approach, a constant cross-sectional area (the two-dimensional analogue of the volumetric sedimentation rate) is iteratively fit to the observed geometry of a half-graben (fig. 4). Throughout the "filling" process, both the half-graben's geometry and the stratal geometry are maintained. As seen in figures 4B and 4C, both the cumulative thickness and accumulation rate curves for both a drill hole through the deepest part of the basin and the average thickness of each stratal package are quite similar in form to the trends produced in a full-graben (cf. figs. 4 and 9), implying that full- and half-graben fill similarly.

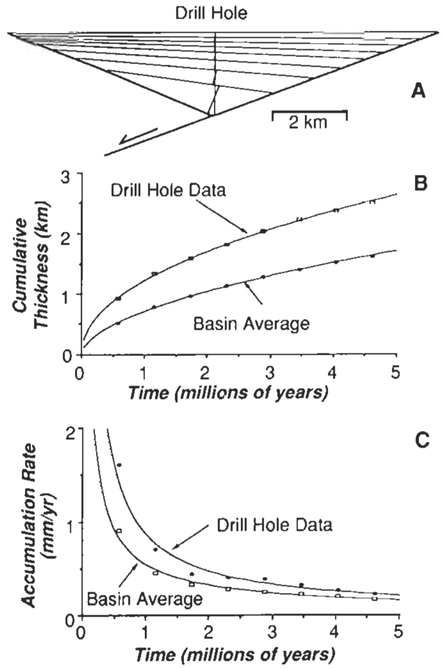

FIG. 4.—Half-graben filling model. (A) Cross section of half-graben loosely based on the geometry of the Atlantis basin (fig. 1), in which stratal surfaces, corresponding to time slices of 550,000 yrs, delineate areas of 1.5×10^6 m^2 which were iteratively fit to the basin. Cumulative sediment thickness (B) and accumulation rate (C) show similar trends as in a full-graben (cf. fig. 9).

The evolution of half-graben, however, can be more complicated than that proposed in the model. The depositional surface area of a half-graben ceases to grow when the onlapping edge of the basin fill passes inside the margin of previously deposited sediments, i.e., when strata cease to onlap the "basement" rocks and begin to onlap previously deposited strata. At that point, under conditions of uniform subsidence, the accumulation rate would cease to change or actually decrease. Additional complications arise in half-graben bounded by domino-style rotational faults. With continued extension, the normal faults progressively rotate to shallower dips (Wernicke and Burchfiel 1982). Even under conditions of uniform extension, therefore, subsidence will not be uniform. Furthermore, subsidence can be expected to decrease from a maximum along the boundary fault to a minimum at the hinged margin of the half-graben. We also expect significant along-strike changes in subsidence, ranging from a maximum near the center of the boundary fault to zero at its tips. This cumulative variation in subsidence would then mimic the nature of subsidence observed following slip on normal faults, such as the 1959 Hebgen Lake earthquake in Montana (Fraser et al. 1964).

The qualitative model presented above predicts major changes in depositional environments as a consequence of the filling of the basin: fluvial sedimentation (basin is sedimentologically and hydrologically open) giving way to hydrologically open lacustrine sedimentation followed by hydrologically closed lacustrine sedimentation. Furthermore, the model can be quantified to predict the timing of the major transitions in depositional environments and trends in accumulation rates and lake depth. Clearly, however, the model cannot address all aspects of extensional basin stratigraphy. For example, during "lacustrine deposition," fluvial and fluvio-deltaic deposits will ring the lacustrine deposits (see facies models of Leeder and Gawthorpe 1987). Furthermore, near the main border faults of the basin, lacustrine and alluvial fan deposits will interfinger. Repeated motions along the border fault will lead to the formation of tectonic cyclothems (Blair and Bilodeau 1988). In addition, the facies distribution will be influenced by climatic changes (Olsen 1984*a*, 1984*b*, 1986): the lakes may shrink and swell, and the position of the marginal lacustrine facies will migrate accordingly. The model therefore is only intended as a guide to understanding the gross stratigraphic development of extensional basins, reflecting perhaps those processes operating on a scale of 10^6 or 10^7 years.

The rationale of the assumptions used in the basin filling model warrants additional discussion. The model assumes a constant volumetric sedimentation rate, yet as the basin fills, younger strata progressively onlap potential source areas of new sediment. As a result, volumetric sedimentation rate might decrease. However, in the East African rifts, most sediments are derived from the hanging wall block and from rivers flowing along the axis of the rift system (Lambiase 1990). If this axial component is a general feature of most rift systems, then the effect of onlap onto source regions may not critically affect sediment supply, and the assumption of constant V_s is probably a good starting point.

The model also assumes that the outlet of the depositional basin is fixed with respect to an external reference system. Based again on the East African rifts, the lowest elevation outlet is commonly found at the along-strike end of an individual half-graben (Lambiase 1990). As this is the region that experiences minimal hanging wall subsidence and footwall uplift, then the outlet may indeed be assumed to be stationary.

As for subsidence and volume of water available, there are no reliable, long-term, modern data. Parsimony therefore dictates that we assume these parameters to be constant. The model is specifically designed to assess the intrinsic effects of the filling of the basin. Only when these are understood can we begin to examine and decipher the causes of the extrinsic effects (changes in subsidence rate, climate, and rates of sediment and water input). This approach is illustrated below with reference to the extensional basins containing the rocks of the early Mesozoic-age Newark Supergroup, particularly the Newark basin.

APPLICATION OF THE BASIN FILLING MODEL

Geologic Overview of the Newark Supergroup.—Eastern North America forms part of the classic Atlantic-type passive continental margin (Bally 1981) created by rifting of the supercontinent of Pangaea. All along the axis of the future Atlantic Ocean from Greenland to Mexico, the Triassic initiation of the breakup was marked by the formation of extended crust. Early Mesozoic rift basins of eastern North America are exposed from South Carolina to Nova Scotia; numerous others have been discovered or inferred by

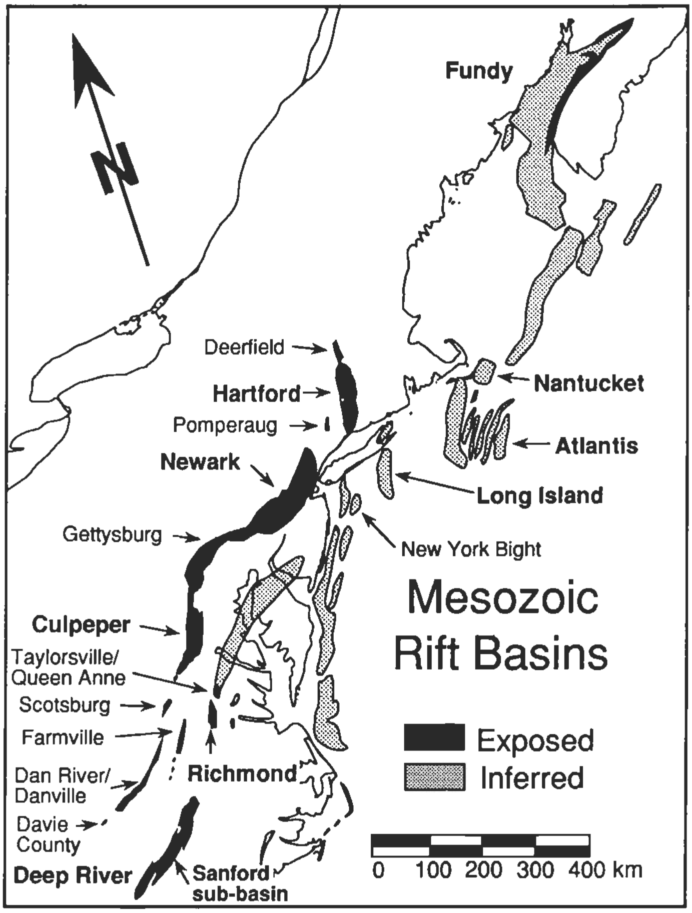

FIG. 5.—Early Mesozoic rift basins of eastern North America. Names of basins mentioned in the text are bold-faced. Modified from Olsen et al. (1989).

drilling, seismic reflection profiling, and gravity studies to be buried both by post-rift sediments on the continental shelf and below coastal plain sediments (fig. 5). Thousands of meters of exclusively continental strata, tholeiitic lava flows, and diabase plutons filled the exposed rift basins. The faulted, tilted, and eroded strata are termed the Newark Supergroup (Van Houten 1977; Olsen 1978; Froelich and Olsen 1984).

Milankovitch-Period Lacustrine Cycles.—The lacustrine rocks of Newark Supergroup basins show a common theme of repetitive cycles named in honor of their discover, F. B. Van Houten, by Olsen (1986). The fundamental Van Houten cycle consists of three lithologically-distinct divisions which are interpreted as (1) lake transgression, (2) high stand, and (3) regression plus low-stand facies (Olsen 1986). These cycles were apparently produced by the basin-wide rise and fall of water level of very large lakes (Van Houten 1964; Olsen 1984*a*, 1984*b*, 1986; Olsen et al. 1989). Fourier analysis of stratigraphic sections has shown that the Van Houten cycles have a periodicity of approximately 21,000 yrs (Olsen 1986). Furthermore, the 21,000-yr-long cycles are hierarchically arranged in compound cycles of 100,000 and 400,000 yrs. It is this hierarchy that provides the best evidence that lake levels were climatically driven by orbital variation according to the Milankovitch theory (Olsen 1986). These cycles permit well-constrained estimates of accumulation rate (thickness of cycle divided by its duration) in the lacustrine

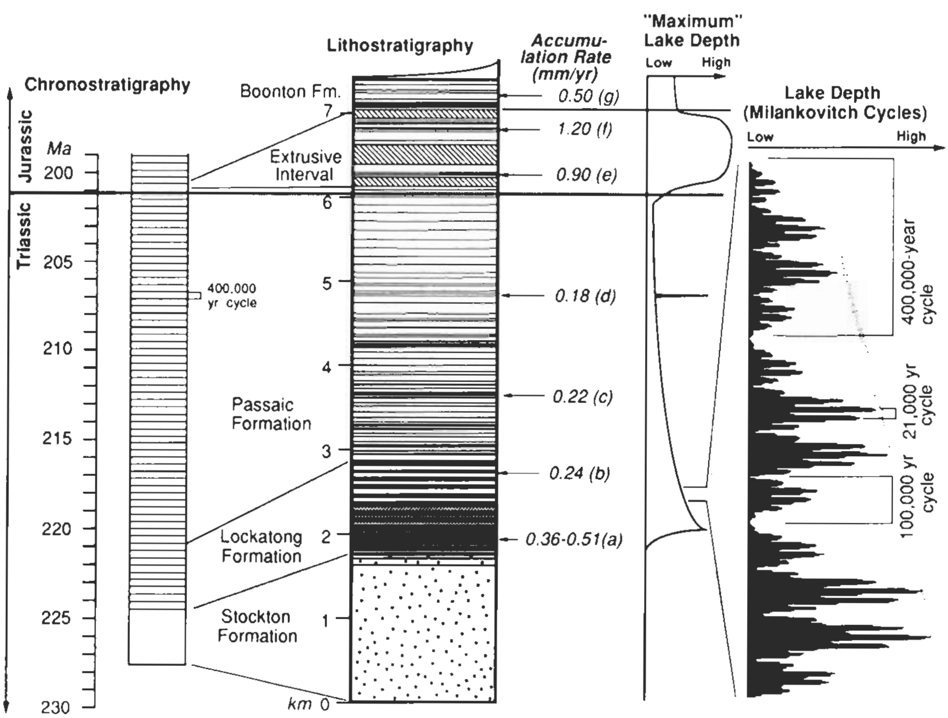

FIG. 6.—Stratigraphic pattern of the Newark basin. Black lines in the stratigraphic column represent the most prominent gray and black units for which their proper stratigraphic position is known. Accumulation rates were calculated using the thicknesses of the 21,000-yr-long Van Houten cycles. Letters refer to specific measured sections: *a*, Lower Lockatong Formation, Lumberville, PA (Olsen unpubl. data); *b*, Middle Lockatong Formation along NJ 29, near Byram, NJ (Olsen 1986); *c*, Perkasie Member of the Passaic Formation, US 202, Dilts Corner, NJ (Olsen et al. 1989); *d*; Ukrainian Member of the Passaic Formation, Ukrainian Village, NJ (Olsen unpubl. data); *e*, *f*, and *g*, Feltville, Towaco, and Boonton formations, respectively, from Army Corps of Engineers drill cores, Pompton Plains to Clifton, NJ (Fedosh and Smoot 1988; Olsen and Fedosh 1988; Olsen et al. 1989). "Maximum" lake depth is the enveloping surface of the Milankovitch-period fluctuations in lake level. The extrusive interval consists of the following formations (in stratigraphic order): Orange Mountain Basalt, Feltville Formation, Preakness Basalt, Towaco Formation, and Hook Mountain Basalt.

intervals of these extensional basins, thereby providing a high-resolution test of the predictions of the basin filling model.

Stratigraphic Pattern.—Two markedly different stratigraphic patterns are present within the basins of the Newark Supergroup and are exemplified by the stratigraphy of the Newark basin (Olsen 1980). The first stratigraphic pattern is of mostly Late Triassic age and consists of (1) the basal red and brown fluvial strata of the Stockton Formation (deposited during the first 3 m.y. of basin subsidence), overlain by (2) the mostly gray and black lacustrine rocks of the Lockatong Formation, which is in turn overlain by (3) the mostly red, lacustrine and marginal lacustrine strata of the Passaic Formation (fig. 6). After the onset of lacustrine deposition, accumulation rate (as calibrated by the thickness of 21,000-yr-long Van Houten cycles) slowly and systematically decreased through the remainder of the Triassic section (Lockatong and Passaic formations). Lake depths were controlled by orbitally induced climatic changes, and the lakes dried out every 21,000 yrs (Olsen 1986). However, certain longer-term trends are also apparent. The first microlaminated black shales formed about 800,000 yrs after the onset of lacustrine deposition and must have been deposited in at least 70–100 m of water, below wave base, at the bottom of an anoxic, turbulently stratified lake (Manspeizer and Olsen 1981; Olsen 1984*a*). The best developed microlaminated units, suggesting the deepest lakes, are found in strata deposited 1.4 m.y. after lacustrine deposition began. Higher in the Lockatong Formation, red massive mudstones become more and more abundant. With the exception of the Ukrainian Member of the Passaic Formation and the Early Jurassic-age sedimentary rocks, the last microlaminated unit is present near the base of the Passaic Formation. Gray and black shales are still present within the rest of the Passaic Formation, but both their frequency and extent of development decrease up-section (fig. 6). For ease in modeling, we have defined "maximum" lake depth as the depth of the deepest lake produced by each 400,000 yr climate cycle; it is the bounding curve for the entire fluctuating lake level curve (see fig. 6). "Maximum" lake depth began shallow at the onset of lacustrine deposition, rapidly deepened during the deposition of the lower Lockatong Formation, and then slowly and systematically decreased toward the Triassic-Jurassic boundary (fig. 6).

The second stratigraphic sequence of mostly Early Jurassic age consists of tholeiitic lava flows interbedded with and succeeded by mostly lacustrine strata, which show marked changes in accumulation rate and "maximum" lake depth. Approximately 30 m below the palynologically determined Triassic-Jurassic boundary (Cornet 1977), both the accumulation rate and "maximum" lake depth greatly increased (fig. 6). The highest accumulation rates and the deepest lake facies (well-developed microlaminated units) are associated with the thick extrusive basalt flows. After the extrusion of the last basalt flow, accumulation rate and "maximum" lake depth again decreased systematically.

Basin Geometry.—Eastern North American rift basins were profoundly influenced by pre-existing structural control (Ratcliffe and Burton 1985; Swanson 1986; Schlische and Olsen 1987, 1988). Where the basins trended oblique to the regional extension direction, extensional strike-slip duplexes, flower structures, and small syndepositional horsts and basins were developed in strike-slip zones (Schlische and Olsen 1987; Olsen et al. 1989). In both volume of sedimentary fill and surface area, however, these strike-slip regions are subordinate to the predominantly dip-slip half-graben.

Where the early Mesozoic extension direction was normal to the strike of pre-existing structures, Paleozoic thrust faults of the Appalachian orogen were reactivated principally as dip-slip normal faults, as in the Newark basin (Ratcliffe and Burton 1985), forming half-graben in their collapsed hanging walls. Based on surface geology and seismic reflection profiles, the basin fill of these half-graben is wedge-shaped (see fig. 1), indicating that normal faulting responsible for basin subsidence was active during sedimentation. Furthermore, younger strata dip less steeply than older strata, indicating syndepositional rotation of the hanging wall block. As mentioned above, published seismic reflection profiles of offshore basins clearly show onlap onto the hanging wall, and studies of outcrops, drill holes, and proprietary seismic reflection profiles of the Newark, Fundy, and Richmond basins lead to a similar conclusion.

The Newark basin of New York, New Jersey, and Pennsylvania is a half-graben bounded on its northwestern side by a southeast-dipping border fault system (figs. 1, 7). The dip of the border fault ranges from 60° to less than 25° (Ratcliffe and Burton 1985). The basin is approximately 190 km long (excluding the narrow neck between the Newark and Gettysburg basins), presently 30–50 km wide, and contains cumulatively >7 km of sedimentary rocks, lava flows, and diabase sheets and dikes. Strata within the basin generally dip at less than 20° toward the border fault system, although they are locally more steeply dipping within transverse folds adjacent to the border and intrabasinal faults (see fig. 7). McLaughlin's studies of the Stockton Formation (McLaughlin 1945; McLaughlin and Willard 1949; Willard et al. 1959) and our mapping of the thickness of Van Houten cycles (Olsen et al. 1989; Olsen 1989) clearly show that units of all ages thicken noticeably toward the border fault side of the basin, indicating that the faulting responsible for basin subsidence was occurring throughout the filling of the basin (contra Faill 1973). The basin is cut by a number of intrabasinal faults, but these appear to have developed late in the history of the basin, after deposition of most of the preserved sedimentary rocks (Olsen 1985; Faill 1988; Schlische and Olsen 1988).

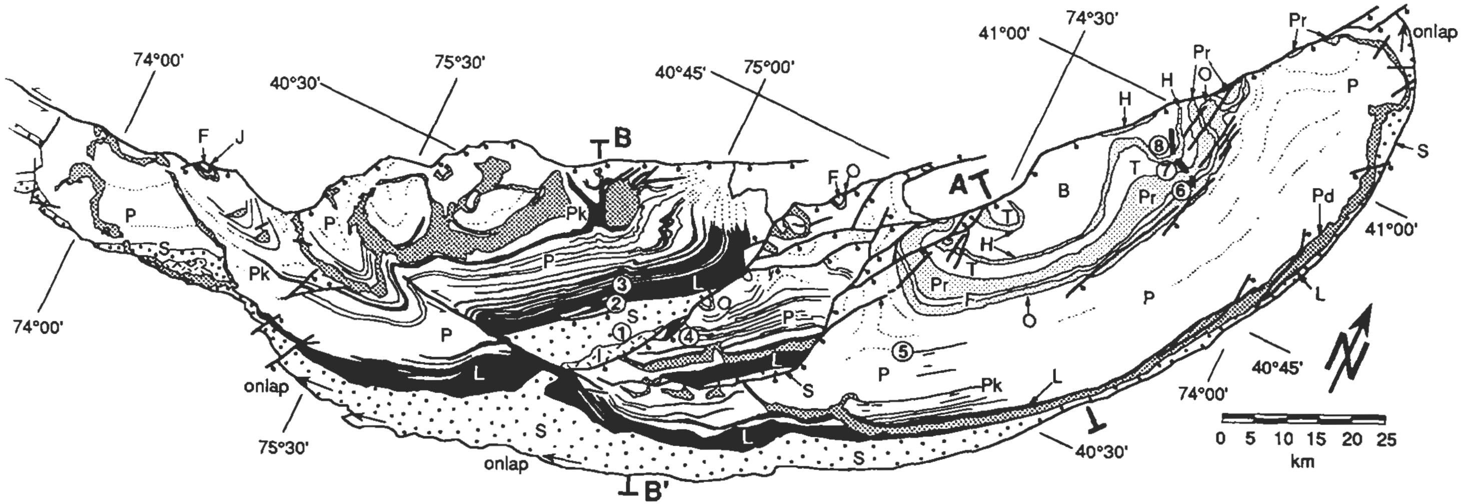

FIG. 7.—Geologic map of the Newark basin of Pennsylvania, New Jersey, and New York. Dark shading represents diabase intrusions, light shading represents lava flows. Dotted lines are from lines of bedding, and thin black lines are gray and black units in the Passaic Formation. Balls are on downthrown side of normal faults. *A* is position of cross section shown in figure 1; cross section B-B′ shown in figure 8*A*. Abbreviations are: *B*, Boonton Formation; *F*, Feltville Formation; *H*, Hook Mountain Basalt; *I*, basement inliers; *J*, Jacksonwald Basalt; *L*, Lockatong Formation; *O*, Orange Mountain Basalt; *P*, Passaic Formation; *Pd*, Palisades diabase; *Pk*, Perkasie Member of Passaic Formation; *Pr*, Preakness Basalt; *S*, Stockton Formation; and *T*, Towaco Formation. Numbers refer to specific sections discussed in text and figure 9: *1*, Stockton, NJ; *2*, Lumberville, PA; *3*, Byram, NJ; *4*, Dilts Corner, NJ; *5*, Ukrainian Village, NJ; *6*, *7*, *8*, Army Corps of Engineers cores, Pompton Lakes to Clifton, NJ. Modified from Schlische and Olsen (1988).

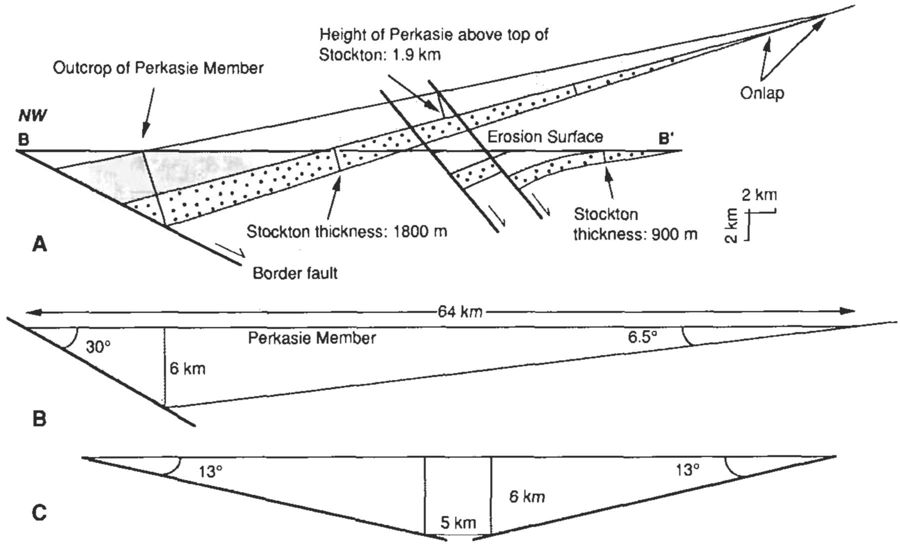

FIG. 8.—(*A*) Cross section A-A′ of the Delaware River valley (see fig. 7 for location) before and after palinspastic restoration of slip on the intrabasinal faults. (*B*) Cross-sectional geometry of palinspastically-restored basin just after the deposition of the Perkasie Member. (*C*) Full-graben equivalent of basin shown in (*B*). The areas of the shaded regions in (*B*) and (*C*) are equal.

Comparison of Model Predictions with Newark Basin Stratigraphy.—Before the full-graben model can be applied, the asymmetric Newark half-graben needs to be "transformed" to a symmetric basin. The approach is to determine a representative cross-sectional area of the Newark basin for a given time-slice and then by conserving cross-sectional area convert the basin into a full-graben with border faults dipping at equal angles. Thus the width of the basin after deposition of a certain stratigraphic horizon and the depth of the basin for that time need to be known. Unfortunately, the Newark basin has been substantially eroded, and its present boundaries do not reflect its limits during deposition. Variations in thickness of units with distance from the border fault, however, do provide important constraints. In the Delaware River valley, the Newark basin is cut up into three fault blocks with repetition of the same formations (fig. 8*A*). Outcrops in the northwestern fault block indicate that the Stockton Formation is approximately 1800 m thick (Olsen 1980), whereas the Stockton is only about 900 m thick where it outcrops in the southeastern fault block (Willard et al. 1959). Because the intrabasinal faulting completely post-dates the preserved strata in these fault blocks (Olsen 1985; Faill 1988; Schlische and Olsen 1988), key horizons can be palinspastically restored as shown in figure 8*A*. In this restored basin, the base and top of the Stockton Formation are constrained by known thicknesses and by the average dip of the base of the Stockton in the northwestern fault block; where these two horizons converge is taken to be the southeastern edge of the Newark basin after deposition of the Stockton Formation. A similar exercise can be performed on the Perkasie Member, where the constraints are (1) the position of Perkasie outcrops in the northwestern fault block and (2) the height of the Perkasie above the top of the Stockton Formation in the middle fault block. A projection of the line connecting these two points clearly extends further to the southeast than the limits of the Stockton Formation, again reflecting the progressive onlap of younger strata onto basement on the hanging wall margin of the basin.

In figure 8*B*, the horizon representing the Perkasie Member has been rotated to paleohorizontal. The basin was approximately 6 km deep in its depocenter (corroborated by proprietary seismic lines) and >60 km wide after deposition of the Perkasie Member. We have assumed that the border fault did not rotate along with the hanging wall and hence the 30° dip shown in figure 8*B* is the same as the present day dip (Ratcliffe et al. 1986). The "full-graben" equivalent of this basin is shown in figure 8*C*. By preserving cross-sectional area, choosing an initial border fault spacing of 5 km (so that the "full-graben" more closely approximates the triangular cross-sectional geometry of the palinspastically-restored Newark half-graben), and by preserving the same 6 km thickness of strata, the "border faults" of the "full-graben" must dip at approximately 13°. Because of this transformation, the output of the model (always given for the deepest part of the full-graben) will most closely match the stratigraphic data from the deepest portion of the Newark half-graben.

Additional parameters entered into the basin filling model were determined as follows. The subsidence rate can be constrained because, under fluvial sedimentation during deposition of the Stockton Formation, subsidence rate equaled accumulation rate. The maximum thickness of the Stockton Formation in the deepest part of the basin is estimated to be 2500 m (see fig. 8*A*), and its paleontologically estimated duration is 3 m.y. (Cornet and Olsen 1985), thus yielding an estimated accumulation rate of 0.833 mm/yr. A volumetric sedimentation rate ($V_s = 4.15 \times 10^6$ m^3/yr) was chosen to yield a transition from fluvial to lacustrine deposition 3 m.y. after the start of subsidence. The maximum volume of water available ($V_w = 1.10 \times 10^{12}$ m^2) was chosen to yield the deepest lakes in the lower Lockatong Formation.

It must be emphasized again that the Newark basin did not evolve as a full-graben. This exercise is only a heuristic procedure meant to demonstrate an application of the basin filling model, not to precisely model the evolution of the Newark basin. It is clear that the approach has severe limitations: namely that it cannot predict transverse and along-strike changes in accumulation rate and lake depth in an asymmetrically-subsiding basin. Nonetheless, the trends in accumulation rate and "maximum" lake depth generated by the model in many instances match the same trends observed in the rock record.

The curves generated using equations (1–3) are shown in figure 9 with reference to actual stratigraphic data. Using the parameters outlined above, the model can generate a transition from fluvial to lacustrine sedimentation at the appropriate time (fig. 9: *b*, *b*′, and *b*″). It remains to be seen whether the V_s needed to generate this transition is realistic. A potential test would be to calculate the volume of a key marker bed, such as the Perkasie Member, over the entire basin; unfortunately, the trends in thickness of units along-strike are not yet well enough defined. The model also predicts a systematic decrease in accumulation rate in the Lockatong and Passaic formations (fig. 9:*f*). Actual stratigraphic data show a similar trend, suggesting that the filling of a basin growing in size through time might indeed be responsible. Notice, however, that the model's accumulation rates are consistently higher than those measured in the basin. If we assume that the sediment added to the model basin was fully compacted, then the discrepancy is most likely the result of the fact that the model's output applies to the deepest part of the basin and that the actual stratigraphic data were not collected from its depocenter. Additional comparisons and extrapolations are not warranted at this time because the stratigraphic data come from outcrops scattered throughout a basin in which appreciable transverse and along-strike changes occur.

The decreasing accumulation rates in the Lockatong and Passaic formations should not be necessarily construed to reflect decreasing subsidence rates. It can be argued that the lake cycles present in these formations represent filling sequences: the basin subsided, formed a lake, which subsequently filled in, the basin subsided, etc. In this scenario, the decompacted cycle thickness would approximate the deepest lake and the amount of subsidence. This argument is flawed on two accounts: (1) The cycle thickness in the Lockatong is only 5.5 m, decompacted to about 10 m, implying a lake depth of 10 m, yet the facies analysis of Olsen (1984*a*) requires the lakes that deposited microlaminated sediments were minimally 70–100 m deep. (2) Most of the Passaic Formation is not microlaminated, and therefore there are no good constraints on lake depth; it is then plausible that subsidence approximates decompacted cycle thickness and was decreasing through the deposition of the Passaic Formation. However, the Ukrainian Member in the middle Passaic is microlaminated and

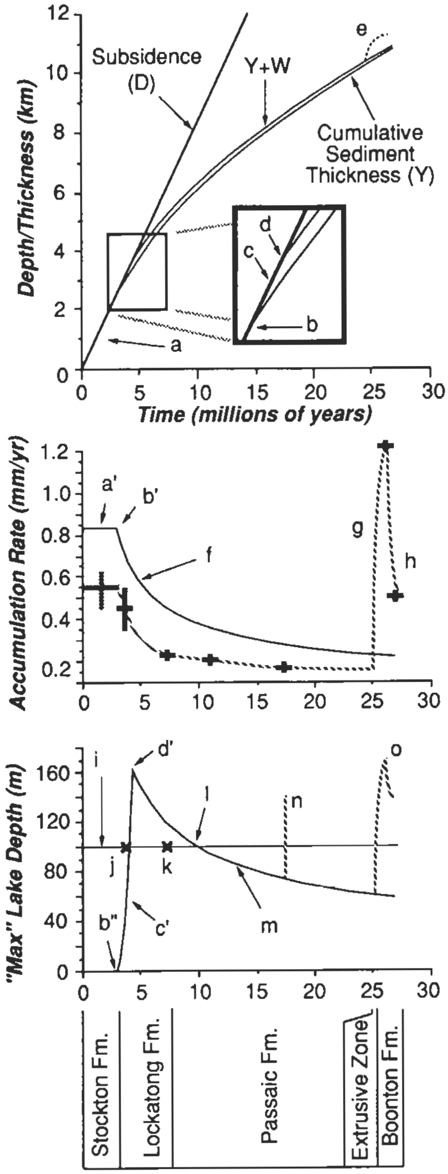

FIG. 9.—Predictions of the full-graben model (solid lines), based on the geometry depicted in figs. 7 and 8, compared to data from the Newark basin (dotted lines and crosses). The $Y + W$ curve represents the height of the lake above the floor of the graben. Abbreviations are: *a*, filling equals subsidence resulting in fluvial sedimentation, with accumulation rate (a') equal to subsidence rate; *b*, b', and b'', onset of lacustrine deposition and sedimentologic closure; *c* and c', rapid increase in lake depth as depositional surface subsides below hydrologic outlet; *d* and d', deepest lake predicted by model and onset of hydrologic closure; *e*, major deviation from cumulative thickness curve in Early Jurassic; *f*, hyperbolic decrease in accumulation rate predicted by model; solid cross represents data on accumulation rates from cyclical lacustrine strata from scattered outcrops in the basin (see fig. 7); shaded cross represents estimate of accumulation rate in fluvial Stockton [estimated thickness (calculated from outcrops near location *1*, fig. 7) divided by estimated duration]; *g*, anomalously high accumulation rate in extrusive zone, a deviation from the model's predictions; *h*, decreasing accumulation rate in the Boonton Formation; *i*, 100 m water depth line, the minimum depth for the formation of microlaminated sediments (Olsen 1984*a*); *j*, actual first occurrence of microlaminated sediments; *k*, actual last occurrence of microlaminated sediments in lower Passaic Formation; *l*, predicted last occurrence of microlaminated sediments; *m*, slow decrease in lake depth; *n*, anomalous "superwet" climatic anomaly of the Ukrainian Member; *o*, anomalously deep Jurassic lakes.

would require an instantaneous 70–100 m of subsidence, which had no effect on accumulation rate (see below). If subsidence was relatively uniform as predicted by the basin filling model, the basin would have been sediment-starved during Passaic deposition, and the basin need not have undergone a radical deepening to accommodate the Ukrainian lake. All we can say for certain is that the subsidence rate was greater than the accumulation rate.

The basin filling model also predicts the very rapid rise in "maximum" lake depth inferred from the lower Lockatong Formation and its subsequent decline (fig. 9:*m*). It is also plausible that longer-term changes in climate were responsible for the trends in "maximum" lake depth. Clearly, climatic changes played a large role in the basin's microstratigraphy (Olsen 1986). However, there is no evidence that the trends in accumulation rate were governed by changes in climate. For example, in the upper Passaic Formation, the range of facies is identical to those in the *driest* portions of the Lockatong Formation (Olsen 1984*a*), yet the accumulation rates in the Passaic Formation are lower. Furthermore, in the rare instances where deep lakes were produced in the Passaic Formation (i.e., the Ukrainian Member), there is no appreciable change in accumulation rate from the surrounding dry facies (see fig. 9). Since trends in accumulation rate appear to be detached from a climatic origin, it seems reasonable (and parsimonious) to assume that changes in "maximum" lake depth might be similarly detached, although to a lesser extent. Note also that the model predicts that the last microlaminated black shale should form considerably later than is indicated by the data (fig. 9:*k*). This most likely can be attributed to the fact that the model did not incorporate the effects of evaporation of water from a lake whose surface area was continually increasing. Thus, actual lake depth decreased more rapidly than predicted by the model.

Operating under its basic assumptions (uniform subsidence, constant rates of input of sediment and water), the model *cannot* explain the observed Early Jurassic increases in *both* accumulation rate and "maximum" lake depth (fig. 9:*g*, *o*), which imply that the Newark basin had actually decreased in size. In full-graben, changes in subsidence have either no effect or the wrong effect, for accumulation rate and "maximum" lake depth are governed by the areas of the depositional and lake surfaces respectively. Increasing the subsidence rate only deepens the basin, but it cannot change the surface areas. Decreasing the subsidence rate initially has no effect until the basin fills to its lowest outlet, whereupon fluvial sedimentation returns.

In half-graben, however, maximum subsidence occurs adjacent to the border fault and decreases systematically toward the basin's hinge. An increase in the subsidence rate results in an increase in basin asymmetry. As a result, sediments and water shift to the deepest portion of the basin. A rapid increase in basin asymmetry in the earliest Jurassic of the Newark basin therefore could have resulted in (temporary) decreases in the areas of the depositional and lake surfaces, causing increases in accumulation rate and "maximum" lake depth. Systematic decreases in accumulation rate and "maximum" lake depth recorded in the Boonton Formation (figs. 6 and 9:*h*) might then reflect the return to uniform subsidence, as the newly deposited sediments progressively onlapped the older sediments and "basement" rocks of the hanging wall.

Asymmetry may have been introduced by two fundamentally different processes (fig. 10): (1) accelerated rotation of the entire hanging wall block and (2) the formation of antithetic faults between the basin's hinge and the border fault, thus generating a narrower sub-basin. Presently available geologic data are not good enough to distinguish between these two processes.

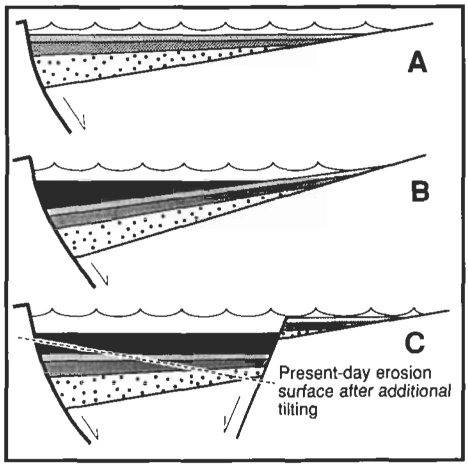

FIG. 10.—Effects of asymmetric basin subsidence. (*A*) Normal filling sequence under conditions of uniform subsidence. (*B*) Accelerated tilting causes sediment and water to shift toward the border fault side of the basin and results in the onlap of youngest strata onto previously-deposited strata. (*C*) Antithetic faulting creates a smaller sub-basin. Fault need not be located within present extent of the basin. Darkest shading in (*B*) and (*C*) corresponds to Jurassic deposits of the Newark basin.

A dramatic increase in the volume of sediment added to the basin per unit time also would have increased the accumulation rate in the Jurassic section. However, this added volume of sediment would have raised the depositional surface and thereby displaced the volume of water upward, perhaps even causing lake level to reach the hydrologic outlet. In a concave-upward basin, this would have had the effect of increasing the lake's surface area, thereby *decreasing* "maximum" lake depth, not increasing it dramatically. An increase in the volume of water (V_w) added to the basin during the wettest portions of the 400,000 yr climate cycles would have increased "maximum" lake depth and presumably also would have brought in more sediments, increasing the accumulation rate. However, this would imply that the accumulation rate ought to be lower in the drier portions of the 400,000 yr cycles, which is not the case. It therefore appears likely that the observed Early Jurassic increases in accumulation rate and "maximum" lake depth resulted from an increase in basin asymmetry.

In terms of the approach taken in this paper, only one parameter—subsidence rate—

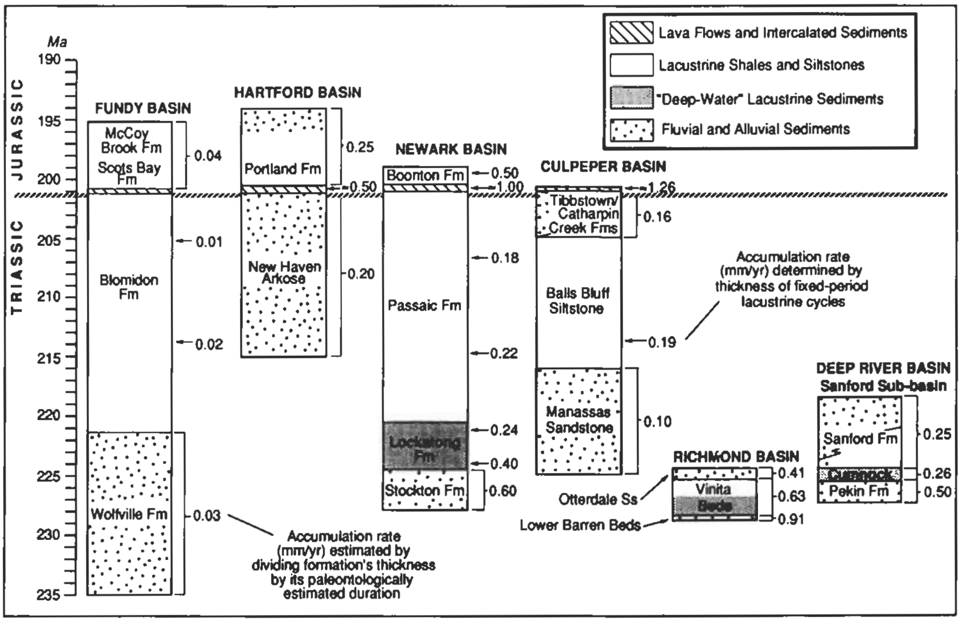

FIG. 11.—Chronostratigraphic sections of several Newark Supergroup basins, showing trends in accumulation rate. In all of the basins, the transition from fluvial to lacustrine deposition occurred at different times. The marked increases in accumulation rate and "maximum" lake depth associated with the extrusive episode occurred simultaneously in all basins that contain a Jurassic section. Modified from Olsen et al. (1989).

needs to be altered to explain the marked changes in the Jurassic sequence. In terms of the traditional approach, the changes would be explained by increasing both the volume of sediment and water added to the basin. For example, Manspeizer (1988) cites the deep Jurassic lakes as evidence of a wetter Jurassic climate. Our new approach indicates that climate need not necessarily have been responsible.

Other Basins.—The basin filling model predicts a basic tripartite stratigraphic pattern consisting of (1) a fluvial base, which would not be present if the subsidence rate is high and/or the rate of sediment input is low), (2) a middle lacustrine interval showing decreasing accumulation rates and evidence of deep lakes near its base followed by gradual shoaling, and (3) possibly an upper fluvial interval, deposited during waning subsidence. These predictions provide a convenient frame of reference against which the observed stratigraphy of basins can be compared and their histories deciphered.

All of the Newark Supergroup basins shown in figure 11 contain a basal fluvial unit (Reinemund 1955; Hubert et al. 1978; Hubert and Forlenza 1988; Ressetar and Taylor 1988; Olsen et al. 1989; Smoot 1989). In the Fundy, Culpeper, Richmond, and Deep River basins, the fluvial units are overlain by predominantly lacustrine deposits (Reinemund 1955; Ressetar and Taylor 1988; Smoot 1989) with the deepest water intervals definitely occurring near the base of the lacustrine sequence in the Deep River, Richmond, and Fundy basins (Olsen et al. 1989). In the Blomidon Formation of the Fundy basin, studies of scattered outcrops support a decrease in accumulation rate up-section (Olsen et al. 1989). In the Hartford basin, the basal New Haven Arkose is fluvial until just prior to the eruption of the first basalt flow (Olsen et al. 1989). It is possible that the basin was subsiding too slowly or the volumetric sedimentation rate was too high to have permitted the onset of lacustrine deposition. It is also possible that the preserved fluvial strata are marginal to lacustrine strata. Since paleocurrents flowed from east to west at the time of New Haven deposition (Hubert et al. 1978) and the strata now dip to the east, any lacustrine strata undoubtedly now would be eroded. In the upper Triassic portions of the Culpeper and Deep River basins, a higher content of coarse clastic material suggests that fluvial sedimentation may have returned (Olsen et al. 1989), signaling a decrease in the subsidence rate. It also is possible that the coarse clastics reflect the proximity to the border fault or that they are the marginal facies to lacustrine strata deposited in a deeper portion of the basin. More study is needed to resolve this problem. The Otterdale Sandstone of the Richmond basin is largely fluvial (Ressetar and Taylor 1988) and contains evidence of Late Triassic paleo-canyon systems apparently cut by rivers after subsidence had stopped or during uplift of the basin (B. Cornet pers. comm. 1989).

In the Fundy, Hartford, and Culpeper basins, the syn- and post-extrusive formations record higher accumulation rates and greater "maximum" lake depths over the preceding Triassic portions (fig. 11; Olsen et al. 1989), possibly reflecting increased asymmetry in each of these basins in the earliest Jurassic. The uppermost Portland Formation of the Hartford basin appears to be fluvial in origin (Olsen et al. 1989). If no lacustrine strata are located downdip of the exposed fluvial strata, then this interval recorded a slowing of basin subsidence, which allowed the Hartford basin to once again fill its lowest outlet.

Lake Bogoria in the Gregory Rift and the hydrologically-open Lake Tanganyika represent two modern examples (Lambiase and Rodgers 1988; Lambiase 1990) of graben in the first two stages, respectively, of the model's tripartite evolution. The Keweenawan trough of Michigan and Wisconsin (Elmore and Engel 1988) also shows a tripartite stratigraphic architecture.

Jones (1988) presented a graph of cumulative sediment and lava flow thickness plotted against time for the Kenya Rift east of the Elgeyo escarpment. Although there is considerable scatter, the rate of increase of cumulative thickness decreases up-section, which may indicate that the basin was increasing in size through time. A paleomagnetically and radiometrically calibrated plot of cumulative sediment thickness against time for the Ngorora Formation of the Kenya Rift shows the same trend (Tauxe et al. 1985; their fig. 4). The higher accumulation rates in the lower part of the section were interpreted to reflect the rapid infilling of a topographic depression and the subsequent lower accumulation rates to reflect sedimentation keeping up with subsidence, implying that excess sediments were leaving the basin. However, the stratigraphy of the upper two-thirds of the Ngorora Formation is predominantly lacustrine (Bishop and Pickford 1972), implying sedimentologic closure. Thus, the decrease in accumulation rate up-section may be an indication of a basin growing in size through time, fitting our model's predictions.

Cumulative sediment thickness and accumulation rate curves for the Cenozoic Vallecito-Fish Creek basin in the Imperial Valley in California (Johnson et al. 1983; their figs. 4 and 5) bear a strong resemblance to those in figure 9. As the sediments were deposited largely in shallow marine waters, implying that sedimentation kept up with subsidence, Johnson et al. (1983) interpreted the curves to reflect decreasing rates of tectonic subsidence. Alternatively, the basin may have been subsiding uniformly and growing in size through time.

SUMMARY AND DISCUSSION

Stratal onlap geometry in extensional basins indicates that the basins were growing in size through time, and therefore the depositional surface area was also increasing through time. Under conditions of uniform subsidence and constant rate of sediment and water input, a simple basin filling model predicts an initial period of fluvial sedimentation as the volume of sediments available exceeds the capacity of the basin, such that the accumulation rate equals the subsidence rate. As the given volume of sediment is spread over a larger and larger basin volume, a point is reached when the sediments exactly fill the basin; subsequently, lacustrine deposition occurs. Accumulation rate decreases as the sediments are spread over a larger depositional surface area. After the onset of lacustrine deposition, lake depth should rapidly deepen to a maximum (as the finite volume of water available just fills the basin between the hydrologic outlet and the depositional surface) and then systematically decrease as the volume of water is spread over a larger area. If the subsidence rate is fast enough or the sediment input rate slow enough, lacustrine deposition will occur from the outset.

The basin filling model provides a framework for understanding the stratigraphic development of extensional basins, particularly the Newark basin of eastern North America. Many of the observed changes through the stratigraphic sequence can be explained simply as the consequence of the filling of a growing basin: the transition from fluvial to lacustrine deposition in the Triassic-age sequence, the decrease in accumulation rate (calibrated by the thickness of Milankovitch-period lacustrine cycles) after the onset of lacustrine deposition, and the inferred rapid rise and subsequent slower decline in "maximum" lake depth. Other observations require additional explanations. For example, figure 9:*p* shows a marked departure in "maximum" lake depth in the Ukrainian Member of the Passaic Formation. Prior to this type of analysis, the "excursion" would simply have been interpeted as one of the many climatic changes thought to have influenced the basin's stratigraphic development. With the basin filling model and Milankovitch-period modulation accounting for most of the changes in the stratigraphic sequence, this anomaly stands out as something unique—a superwet climatic event. Accumulation rate was not affected, indicating that sediment accumulation rates and climate are not causally linked (fig. 9).

Analogous to thermal modeling of passive margins, the usefulness of the basin filling model rests in subtracting the stratigraphic effects due simply to the filling of the basin (intrinsic effects) from the observed sedimentary record. The resulting anomalies reflect departures from the basin filling model's assumptions (e.g., constant volume of sediment input per unit time, constant inflow rate of water, etc.) and point to important changes in tectonics, climate, and sediment and water budgets (extrinsic effects).

Furthermore, two or more basins may be compared in order to separate local from regional effects. For example, in each of the basins illustrated in figure 11, the transition from fluvial to lacustrine deposition in the Triassic-age sequences occurred at different times (local effect). This is to be expected: even if all the basins had equal subsidence rates and began subsiding simultaneously, their filling rates and the timing of the fluvial-lacustrine transition would depend on the dimensions of the basins, their geometries, and the rates of sediment input.

On the other hand, the marked increases in accumulation rate and "maximum" lake depth occurred simultaneously in the earliest Jurassic in all of the basins containing this sequence (regional effect; fig. 11). This anomaly is attributed to a marked increase in basin asymmetry, causing sediment and water to pile up adjacent to the border fault (fig. 11). This asymmetry may have resulted from a sudden change in border fault geometry. However, this seems unlikely, as it would mean that all border faults experienced a similar change in geometry at the same time. More likely, the basins' asymmetry is attributable to a marked increase in the extension rate, which was probably governed by deep-seated mantle processes.

The inferred regional increase in the extension rate occurred just 150,000 yrs before the volumetrically massive early Mesozoic extrusive event. The sudden increased extension may have thinned the crust sufficiently to allow melting in the upper mantle. That the accelerated extension was a brief excursion is supported by the brief duration (<550,000 yrs) of the eastern North American extrusive episode (Olsen and Fedosh 1988). Furthermore, the loading of up to 700 m of basalt and coeval intrusives within the basins may have contributed to their asymmetry.

Both the filling of a growing extensional basin as a result of mechanical (fault-controlled) subsidence and subsidence induced by decay of thermal anomalies on passive margins produce much the same result on accumulation rates: a decrease in rate up-section. We do not imply that our basin filling model is applicable to deposits resulting from thermal subsidence on known passive margins. However, care must be taken in interpreting accumulation rates in older (particularly fossil-poor) rocks for which precise tectonic and depositional settings (marine vs. continental) may not be known and the cause of the subsidence (mechanical vs. thermal) may be confused. The resolution of this problem may be aided somewhat by the observation that thermal subsidence tends to be regionally persistent, whereas mechanical subsidence generally is restricted to smaller, discrete basins, where diagnostic coarse-grained facies may point to normal faulting as the cause of subsidence.

In basins in which fault-controlled subsidence is thought to be the dominant mechanism of basin evolution, as in the basins of the Newark Supergroup, the minimum subsidence rate can be constrained, for the subsidence rate is greater than the filling rate during lacustrine deposition and equal to the filling rate during fluvial deposition. Ultimately, by combining the results of the basin filling model's subsidence analysis with knowledge on how extensional basins grow and deform through time (e.g., Wernicke and Burchfiel 1982; Gibbs 1984), it may be possible to estimate how the extension rate varied throughout the history of Early Mesozoic extension, provided the basins' geometries are also defined through time.

Before the extensional basin filling model can be used to its full potential, several important improvements need to be made. Foremost, three-dimensional quantitative models for the filling of half-graben need to be developed, models that take into account not only variations in subsidence with position in the basin but also build in the along-strike growth of the basin. The effects of differential relief and the erosion of uplifted blocks on the filling of the basins has yet to be addressed. The non-linear losses of water volume to evaporation still need to be quantified as well as the effects of sediment compaction. Future models also need to incorporate different filling sequences for structurally distinct yet sedimentologically-linked half-graben, such as the Newark-Gettysburg-Culpeper system in eastern North America (Olsen et al. 1989) and the Ruzizi-Kigoma-Southern depositional basins of Lake Tanganyika (Burgess et al. 1988; Lambiase 1990). In addition, the existing geometry of the extensional basins needs to be more tightly constrained, ideally through a network of intersecting transverse and along-strike seismic lines.

Despite the shortcomings of the extensional basin filling model presented in this report, the preliminary results are encouraging: even under the simplest conditions, continental facies in extensional basins are expected to exhibit predictable changes through time as a consequence of the changing geometry of the basin and not necessarily as a result of changing tectonic or climatic parameters. Indeed, in order to recognize changes in the extrinsic parameters, the changes intrinsic to the basin must first be subtracted from the stratigraphic record.

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