Sedimentology (1986) **33**, 487–497

Numerical computation of co-oscillating palaeotides in the Catskill epeiric Sea of eastern North America

RUDY SLINGERLAND

Department of Geosciences, The Pennsylvania State University,
507 Deike Building, University Park, PA 16802, U.S.A.

ABSTRACT

A numerical model describing two-dimensional, shallow water, long wave propagation has provided estimates of the co-oscillating palaeotides in the upper Devonian Catskill Sea. This sea was open to the ocean along the present Gulf Coast and, moving clockwise, was circumscribed by the Transcontinental Arch, the Old Red Sandstone Continent, and the rising Appalachian Orogen. Using the present best estimates of basin bathymetry and open-ocean tidal range, the model indicates high mesotidal to low macrotidal ranges for the Catskill Shelf along the ancestral Appalachians. The Mid-Continent Shelf, centred on Iowa, is considered to have been microtidal. These conclusions are not sensitive to reasonable variations in shelf geometry and support the hypothesis that at least some epicontinental seas could have been tide-dominated.

INTRODUCTION

Throughout Phanerozoic history, large volumes of sediment have been deposited in shallow epeiric and mioclinal seas by processes that still are not well known. One school of thought holds that most of these seas were tideless (Shaw, 1964; Mazzullo & Friedman, 1975) because some modern counterparts, such as Florida Bay, show rapid tidal wave attenuation landward. By default, their important processes presumably were storm(wave)-dominated (Goldring & Langenstrassen, 1979; Dott, 1974). Another school holds that many of these ancient seas were dominated by tides and by tidal circulation patterns (Klein & Ryer, 1978; Klein, 1977) because other modern counterparts, such as the Arafura Sea between Australia and New Guinea, show a positive correlation between tidal range and width. In support of this interpretation, Klein (1977) and Klein & Ryer (1978) reported common tidal indicators in Precambrian, early Palaeozoic, and Mesozoic deposits such as herringbone cross-bedding, reactivation surfaces, wavy and flaser bedding, mud cracks, washouts, 'escape' burrows, and fining-upwards, palaeotidal sequences.

This controversy has now arisen in studies of Late Devonian Catskill rocks of eastern North America. Walker & Harms (1975) concluded that the Irish Valley Member of the Catskill Formation in central Pennsylvania was deposited in a sea of low tidal range because of an absence of channels, a close association of marine and non-marine elements, and an absence of winnowed sand bodies. Similarly, Woodrow & Isley (1983) and Dennison (1985) concluded that the Catskill Sea possessed a tidal range of 1 m or less because it was cut off from the open ocean and present-day, partly enclosed seaways show greatly restricted tidal ranges (Schopf, 1980; Klein & Ryer, 1978). Allen & Friend (1968), however, identified marine cyclothems, the coarse members of which were composed of medium to very fine grained, complexly cross-stratified sandstones, in the same Irish Valley exposures studied by Walker & Harms. They interpreted these as tidal channel deposits (p. 56) by association with those of the North Sea, a macrotidal coast. More recently, Rahmanian (1979, 1980) documented the sub- and inter-tidal character of the Irish Valley Member in outcrops throughout central Pennsylvania. He interpreted large sandstone bodies in western Centre County to be tidal current ridges of a tide-dominated delta that was flanked by less sandy, broad tidal flats. Areally restricted bioclastic carbonate units, such as the Luther's Mills Coquinite in north central Pennsylvania, probably also have a tidal origin. They are composed of polymodal, large-scale cross-bedded, skeletal fragment grainstones with lateral accretion bedding and have been interpreted as subaqueous shell mounds in a tidal flat (Krajewski & Cuffey, 1976), tidal deltas (Woodrow *et al.*, 1981) or subtidal bars in a tide-dominated fluvial delta (Woodrow *et al.*, 1981), and sand bars adjacent to tidal channels (Bridge & Droser, 1986). Others observing tidal deposits are Krajewski & Williams (1971) (tidal bars in the Catskill Formation of north-eastern Pennsylvania) and McCave (1968) (tide-dominated shelf deposits in the upper Ludlowville and Lower Moscow Formations). Thus, two conflicting views have arisen concerning the extent of tidal influence in the Upper Devonian of the central Appalachians.

It is the purpose of this paper to investigate the problem by numerically simulating the dynamic-equilibrium, co-oscillating palaeotides of the Catskill Sea using two-dimensional, long wave equations under appropriate boundary and initial conditions. The rationale is to avoid the weaknesses inherent in comparisons with modern shelves by incorporating as far as possible the actual basin geometry, palaeogeography, and palaeogeophysics of the Catskill Sea. Where input parameters are not well constrained, such as the exact nature of the tides propagating on to the continent from the open ocean, a range of potential values is used. The outcome is a range of water surface elevations and horizontal water velocity vectors at all points of interest throughout a complete tidal cycle. From these data, co-range lines and residual velocity vectors are calculated to define the theoretical tidal regime of the Catskill Sea.

Other theoretical attempts to reconstruct palaeotidal regimes are offered by Sündermann & Brosche (1978), Bridges (1982), Hansen (1982), Krohn & Sündermann (1982), Webb (1982) and Slater (1985). Sündermann & Brosche, Krohn & Sündermann, Hansen and Webb wanted to obtain the oceanic tidal torque through time for their studies on the orbital evolution of the earth–moon–sun system. They solved numerically either the hydrodynamical differential equations or the Laplace tidal equations. By necessity they used an idealized, global ocean from which little can be said concerning tides in epeiric seas. However, they did demonstrate that Palaeozoic ocean tides were of a similar range to those today. Bridges used the dynamics of tidal waves on modern shelves to speculate on the palaeotidal regimes in the Upper Jurassic and Upper Cretaceous epicontinental seaways of western North America. The results for the Upper Cretaceous were refined by Slater who solved the Laplace tidal equations for three different seaway depths and five different combinations of independent and co-oscillating tides. His solutions show appreciable independent tides in the Upper Cretaceous seaway. The present study is undertaken in the same spirit and with the view that enough detail now exists on Upper Devonian palaeogeography in eastern North America to produce realistic hindcasts of palaeotides in the Catskill Sea.

THE TIDAL MODEL

The tidal hydrodynamic model used in this study is modified from Hess & White (1974). The basic equations are the vertically-integrated x and y Navier–Stokes momentum equations and the conservation of mass equation assuming incompressible flow, all in a right-handed coordinate system with the z-axis directed upwards. Important assumptions in their derivation are constant atmospheric pressure, the Boussinesq assumption, uniform fluid density, the Chezy approximation for bottom shear stresses, and no surface stress due to wind. The differential equations for a co-oscillating tide in a shallow ocean are

$$\frac{\partial U}{\partial t}+U\frac{\partial U}{\partial x}+V\frac{\partial U}{\partial y} = -g\frac{\partial \eta}{\partial x}+fV-\frac{gU(U^2+V^2)^{1/2}}{C^2H} \tag{1}$$

$$\frac{\partial V}{\partial t}+U\frac{\partial V}{\partial x}+V\frac{\partial V}{\partial y} = -g\frac{\partial \eta}{\partial y}-fU-\frac{gV(U^2+V^2)^{1/2}}{C^2H} \tag{2}$$

$$\frac{\partial \eta}{\partial t}+\frac{\partial}{\partial x}(HU)+\frac{\partial}{\partial y}(HV)=0 \tag{3}$$

$$H=h+\eta, \tag{4}$$

where U and V are, respectively, the x- and y-directed vertically-averaged water velocities, g is the gravitational acceleration, η is the water surface elevation above or below mean sea-level, f is the Coriolis parameter, C is the Chezy coefficient, h is the distance to the bed below mean sea-level, and t is time. The Devonian value of g is unknown, and so is assumed to be unchanged from the present. The Coriolis parameter is a function of the rate of angular rotation of the earth, the magnitude of which is known to have been larger in the Palaeozoic (Lambeck, 1980). However, it was enhanced in the upper Devonian probably by less than a few per cent (Lambeck, 1980) and its present value is used here.

The effects of friction are introduced by the Chezy coefficient, C, equal to $0{\cdot}816\ H^{1/6}n^{-1}\ \mathrm{m}^{1/2}\ \mathrm{s}^{-1}$ if H is in metres and Manning's n has the units of $\mathrm{m}^{1/6}$. Choosing values of n appropriate for continental shelves is difficult, due to the lack of data; Masch & Brandes (1971) chose values between 0·015 and 0·048 $\mathrm{m}^{1/6}$ for their study of shallow estuaries; Hess & White obtained their most accurate simulation of tides in Narragansett Bay, Rhode Island when n averaged 0·016. For comparison, Rouse (1961) lists values of Manning's n for planed wood (0·008), earth (0·016–0·025) and gravel (0·018–0·029). A value of 0·032 $\mathrm{m}^{1/6}$ was chosen for this study to ensure that any error would be on the side of too much friction. The resulting Chezy values are not unrealistic, and in fact, at the intermediate depths used in this study are similar to those obtained by Dronkers (1964) for the estuary of the River Lek.

As is evident in equations (1–4), this study does not consider the independent tides, that is, tides generated by the direct action of tidal forces on the epeiric body of water. This approach is justified by comparing the Catskill Sea with the Mediterranean, a sea with negligible independent tides. The two are of similar volume, both are located at around 30° latitude, and both are elongate in an east–west direction. Even allowing for differences in bathymetry in the basins, it is probable that co-oscillating tides were dominant over independent tides in the Catskill Sea. This study also does not consider wind set-up or storm surges, both likely to be significant in this epeiric sea.

The method of solution of equations (1–4) is a 'multi-operation' finite difference scheme by Leendertse (1967) which has been encoded into a FORTRAN program by Hess & White (1974) and modified for use in the present study.

APPLICATION TO THE CATSKILL SEA

The Catskill Sea (Woodrow & Isley, 1983) as used here is that epeiric body of water covering eastern North America in Frasnian–Famennian time (Fig. 1).

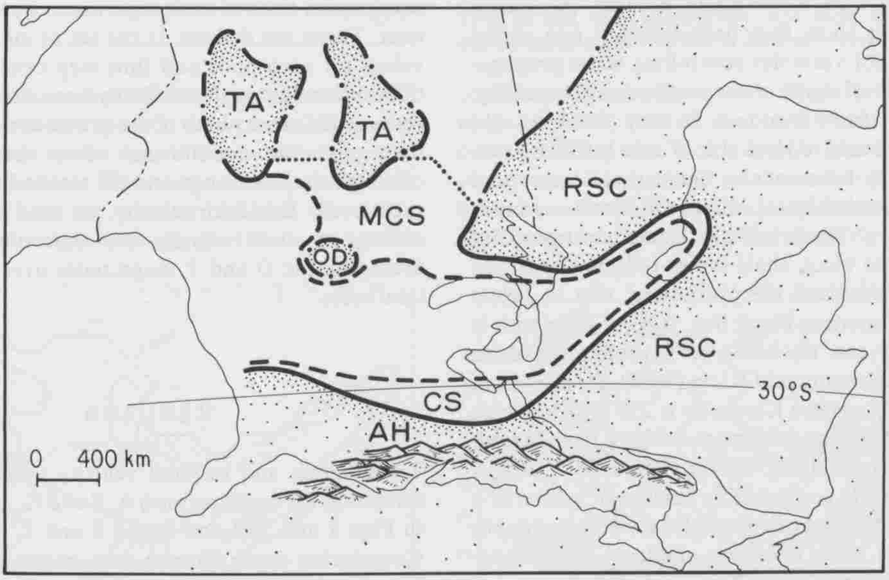


Fig. 1. Palaeogeography of eastern North America in Frasnian–Famennian time adapted from Heckel & Witzke (1979). The dark solid line represents the probable limit of the Catskill Sea on the craton; the dark dot-dashed line represents a less certain limit. The dark dashed line represents the boundary between black shale and predominantly terrigenous detrital facies, taken here as the base of the clinoform. The dotted line represents a boundary to the Catskill Sea defined by shoals. TA = Transcontinental Arch; RSC = Red Sandstone Continent; MCS = Mid-Continent Shelf; OD = Ozark Dome; CS = Catskill Shelf; AH = Appalachian Highlands.

Its palaeogeographic location in the late Devonian is controversial (Boucot & Gray, 1983; Smith, Hurley & Briden, 1981; Ziegler *et al.*, 1979; Heckel & Witzke, 1979; Woodrow, Fletcher & Ahrnsbrak, 1973), ranging from the equator to 30° south latitude in various reconstructions. In this study it is placed between 20 and 30° S latitude following Heckel & Witzke (1979). Choosing a more northerly latitude changes the predicted palaeotidal ranges by less than 5%. The sea opened to the ocean along the present Gulf Coast, and moving clockwise, was circumscribed by the Transcontinental Arch, Old Red Sandstone Continent, and the rising Appalachian Orogen (Heckel & Witzke, 1979). The Ozark Highlands formed an island within it. At various times, the Transcontinental Arch contained straits opening into shallow seas of the present-day Rocky Mountain Cordillera region. Preserved sedimentary facies in the straits indicate very shallow depths however, and to a first approximation the straits are treated as land boundaries in the model. The superimposed finite difference grid (Fig. 2) contains 229 nodes at which η, U and V are calculated, and 11 nodes along the western side at which the open ocean tides must be specified. Node spacings in the x- and y-direction are both equal to 104 km (65 miles).

The bathymetry of the Catskill Sea is moderately well known. This is fortunate because theoretical considerations show that bathymetry is one of the most important variables controlling wave propagation. The overall shape of the seafloor used here (Fig. 2) is adapted from Woodrow & Isley (1983, fig. 3b) and the facies map of Heckel & Witzke (1979, fig. 3c). The boundary between the black shale facies and terrigenous detrital facies defines the clinoform–basin floor boundary. The actual depths are constrained in two ways. The black shale facies reflects deposition below wave base and the pycnocline, that is, below 150 m in the modern Black Sea. Also, if the depth is estimated by the thickness of a prograding delta sequence in the manner of Klein (1974), the minimum basin depth in eastern Kentucky is 230 m (Ettensohn & Barron, 1981); the maximum is 980 m (Lundegard, Samuels & Pryor, 1980; Potter, Maynard & Pryor, 1980). Using the conservative shallower values as a guide, solutions were calculated for six different depth combinations, three of which, shallow (A), intermediate (B) and deep (C), are given in Figs 2–7.

The ocean tide along the western edge of the Catskill Sea, one of the boundary conditions for the model, is not known. In the upper Devonian, the earth–moon distance was a few per cent smaller than its present value (Lambeck, 1980, p. 370) and open-ocean tidal amplitudes should have been larger than those today. Present open-ocean tidal ranges are on the order of 1 m but these increase to an average 2 m on continental shelves and reach over 3 m on nearly 20% of the world's shelves (Schopf, 1980, fig. 3–3). Sündermann & Brosche (1978) computed amplitudes of the M_2-tide in the Pangean Ocean (250–230 My BP) of up to 1·5 m using their best estimates of the earth–moon distance and Coriolis parameter. Given this range, solutions were obtained for six different semidiurnal M_2 tidal waves propagating into the Catskill Sea from the shelf edge with amplitudes ranging from 0.3 to 6 m. Complete solutions are presented here for three boundary tides: 1·8 (microtidal), 3 m (mesotidal), and 6 m (macrotidal) (revised nomenclature of Hayes, 1979). To test the generality of the conclusions, selected solutions were also obtained at intermediate basin depths for a gentler maximum clinoform slope ($2{\cdot}9 \times 10^{-4}$ versus $9{\cdot}6 \times 10^{-4}$), a less concave Catskill shoreline, and a narrower seaway entrance.

Forty-nine numerical experiments in all were conducted using the boundary and initial conditions and constants defined above. Each experiment consisted of calculating η, U and V at each node for as many time steps as it took to eliminate start-up errors caused by inaccurate initial conditions. The dynamic equilibrium tides of each experiment were calculated next. These are defined as the set of all η, U and V values for each node and time step over a complete tidal cycle which do not differ by more than 0·1% from those at the same phase of the previous cycle. Finally these dynamic equilibrium values were used to calculate the tidal range and the residual velocities at each node. Residual velocity, as used here, is the average resultant velocity at a node obtained from averaging the U and V magnitudes over a complete tidal cycle.

RESULTS

Tidal ranges and residual velocity magnitudes and directions for depth regimes A, B and C, are presented in Figs 2 and 3, 4 and 5 and 6 and 7, respectively. Considering depth series A first, co-range lines (Fig. 2) are similar for all three boundary tidal ranges and for each the dynamic equilibrium tidal wave is attenuated rapidly. For example, the 0·5 m co-range line remains in the same location for all three boundary tidal ranges and the eastern (Devonian coordinates) two-thirds of the Sea has at most a 0·5 m tidal range.

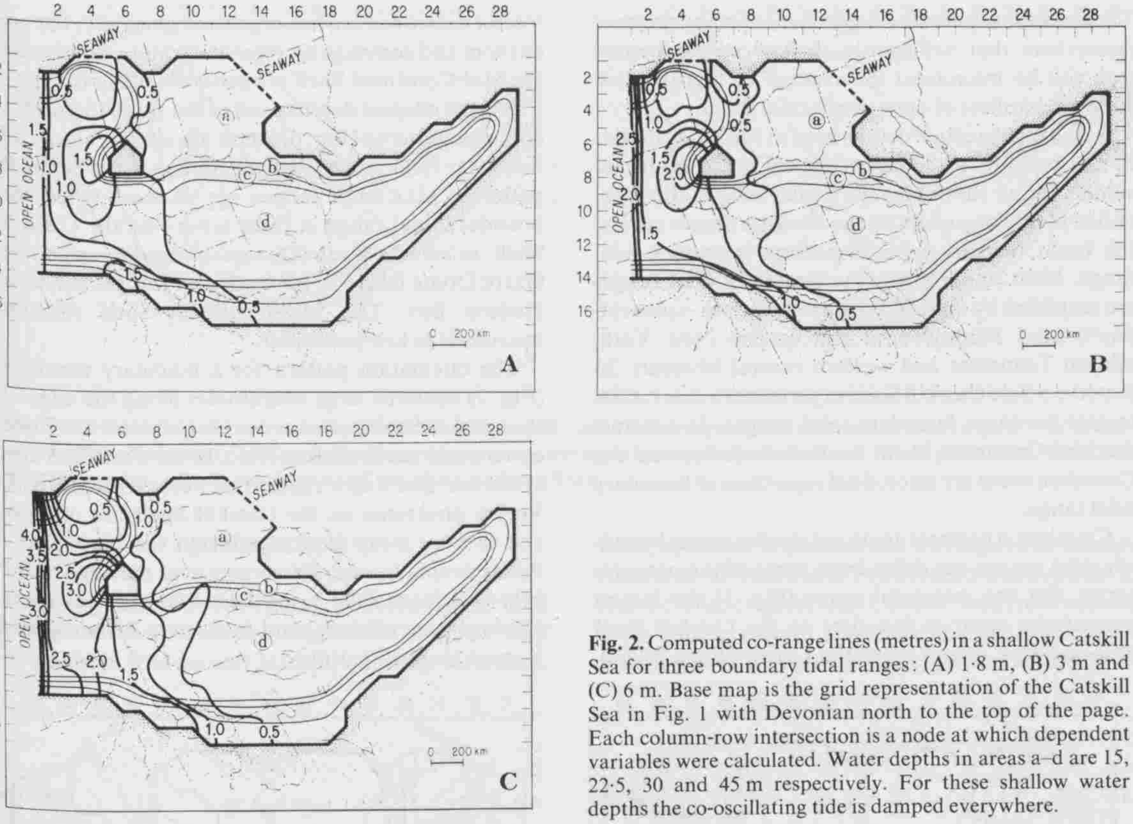


Fig. 2. Computed co-range lines (metres) in a shallow Catskill Sea for three boundary tidal ranges: (A) 1·8 m, (B) 3 m and (C) 6 m. Base map is the grid representation of the Catskill Sea in Fig. 1 with Devonian north to the top of the page. Each column-row intersection is a node at which dependent variables were calculated. Water depths in areas a–d are 15, 22·5, 30 and 45 m respectively. For these shallow water depths the co-oscillating tide is damped everywhere.

The highest tides are less than half the boundary tidal range in each case and occur on the western (Devonian coordinates) side of the Ozark Dome and along the south-western coast. The circulation patterns of residual velocities are also similar for the three boundary tidal ranges and only that for a mesotidal boundary tide is shown here (Fig. 3). The largest magnitudes occur west and south-west (Devonian coordinates) of the Ozark Dome Island. The flow vectors are disorganized on the mid-continent shelf but show a clockwise flow in along the mid-continent shelf slope and out along the Catskill shelf slope.

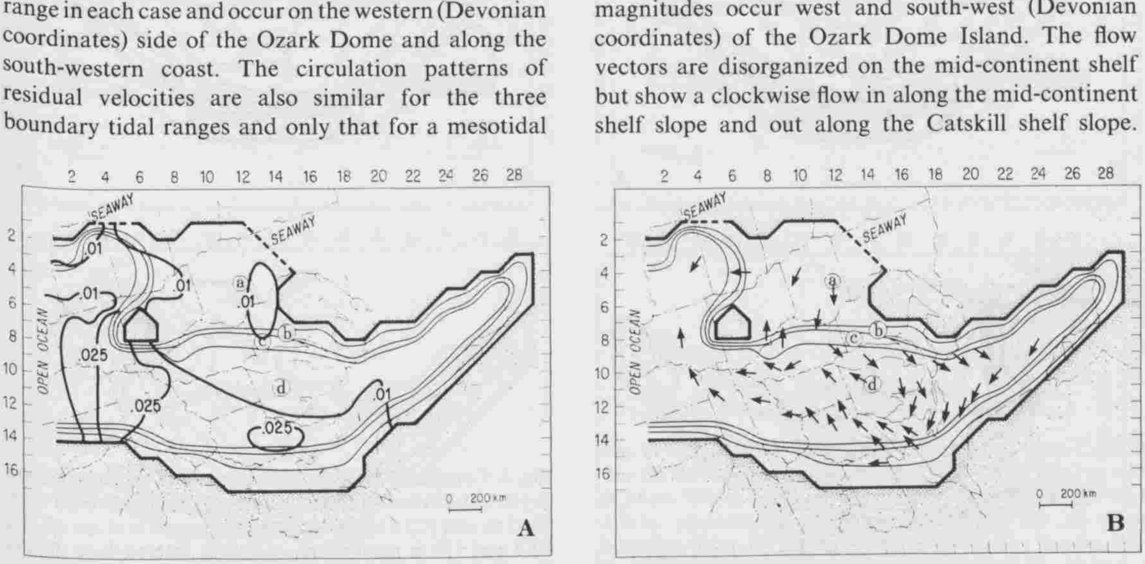


Fig. 3. (A) Residual tidal current magnitudes (m s^{-1}) and (B) directions computed for a boundary tidal range of 3 m and the shallow bathymetry of Fig. 2. Circulation is weak and clockwise over much of the basin.

These results support the suggestions of many previous researchers that sufficiently shallow epicontinental seas will be microtidal for average to high friction factors regardless of open ocean tidal range.

At the intermediate depths used in this study (Series B) the patterns of co-range lines (Fig. 4) again are similar for all three boundary tidal ranges although, unlike at shallower depths, the absolute ranges around the basin increase with increasing boundary tidal range. Most importantly, the boundary tidal ranges are amplified by up to 5/3 in three locations—present-day central Pennsylvania and central New York, eastern Tennessee and western central Missouri. In fact, the whole Catskill Shelf experiences a macrotidal regime for these boundary tidal ranges. In contrast the Mid-Continent Shelf, centred on Iowa, and the Canadian areas are microtidal regardless of boundary tidal range.

Circulation patterns again are similar across boundary tidal ranges but differ from those of the previous series. For the mesotidal range (Fig. 5) the largest magnitudes occur at five sites on the Catskill Shelf and one area inland from the Ozark Dome Island. Vector directions are less organized generally, but are onshore and converge on New York State whereas on the Mid-Continent Shelf are generally offshore.

For the deepest depths used in the study (Series C), co-range patterns (Fig. 6) again are similar across all boundary tidal ranges and are similar to the Series B patterns. Maximum ranges are at least twice the boundary tidal range in three areas—on the Catskill Shelf, in an area north (Devonian coordinates) of the Ozark Dome Island in Missouri and Kansas and near Hudson Bay. The Mid-Continent Shelf remains microtidal to low-mesotidal.

The circulation pattern for a boundary mesotide (Fig. 7) contains large magnitudes along the edge of the Catskill Shelf in Pennsylvania and New York and eastern and south-eastern Ohio, in the Canadian arm of the sea, and a spot centred on west-central Illinois. Vector directions on the Catskill Shelf are offshore and diverge away from a point on the New York–Pennsylvania border. Directions near the edge of the Mid-Continent Shelf in northern Missouri, Iowa and Michigan are offshore, and in western Missouri and Kansas are shelf parallel.

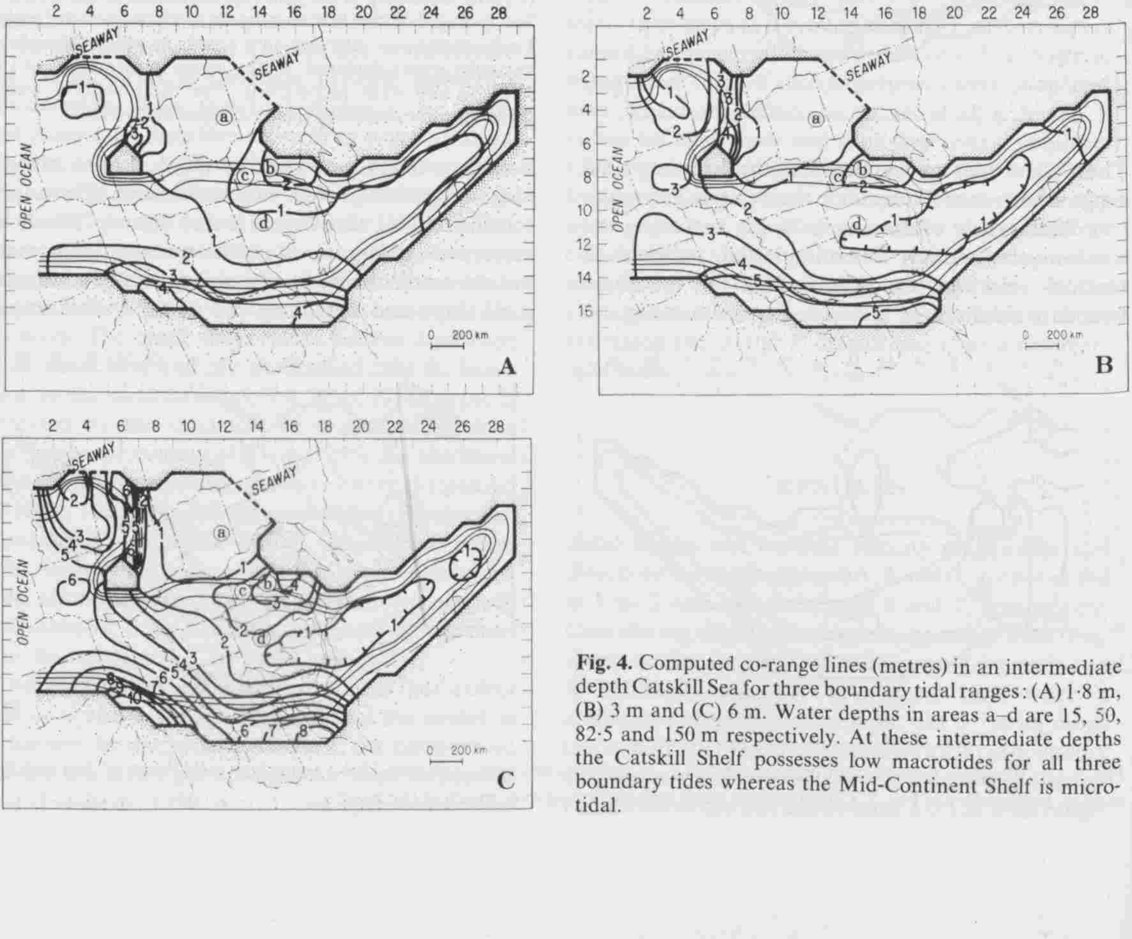


Fig. 4. Computed co-range lines (metres) in an intermediate depth Catskill Sea for three boundary tidal ranges: (A) 1·8 m, (B) 3 m and (C) 6 m. Water depths in areas a–d are 15, 50, 82·5 and 150 m respectively. At these intermediate depths the Catskill Shelf possesses low macrotides for all three boundary tides whereas the Mid-Continent Shelf is microtidal.

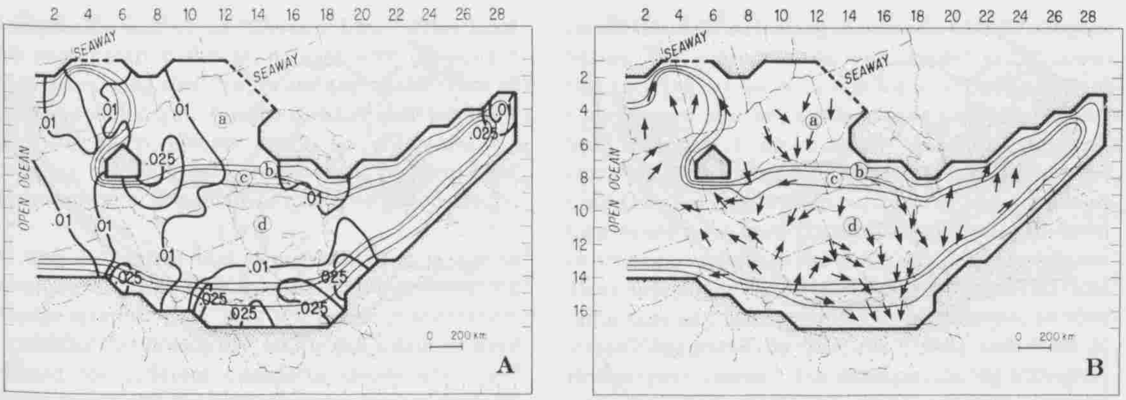
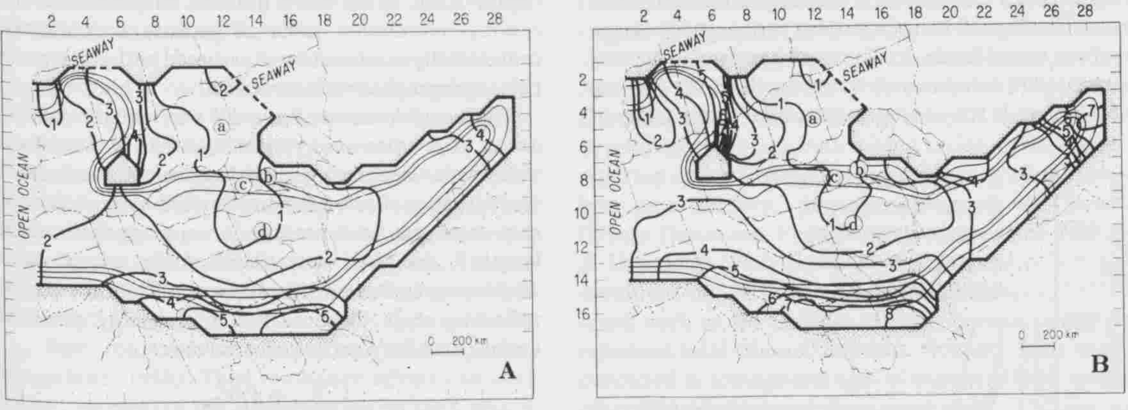


Fig. 5. (A) Residual tidal current magnitudes (m s^{-1}) and (B) directions computed for a boundary tidal range of 3 m and the intermediate bathymetry of Fig. 4. Residual velocities are strongest on the Catskill Shelf and converge on New York State whereas weaker currents are directed off the Mid-Continent Shelf.

DISCUSSION

Certain generalizations emerge concerning the tidal regime of the Catskill Sea. The Mid-Continent Shelf of northern Kansas and Missouri, eastern Nebraska, Iowa, southern Wisconsin and Michigan and western Illinois in all likelihood experienced a microtidal to low-mesotidal regime. The Catskill Shelf from western New York to eastern Tennessee and the west-central Missouri and northern Kansas portion of the Mid-



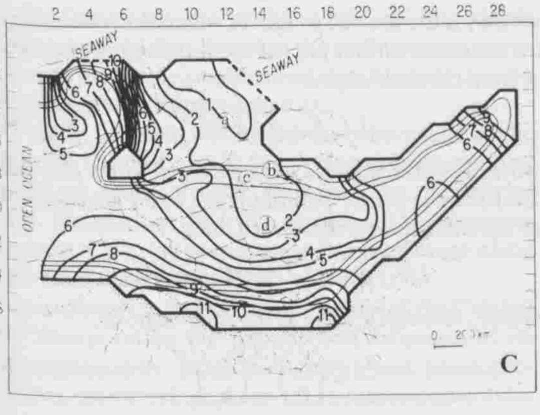


Fig. 6. Computed co-range lines (metres) in a deep Catskill Sea for three boundary tidal ranges: (A) 1·8 m, (B) 3 m and (C) 6 m. Water depths in the areas a–d are 30, 60, 120 and 300 m, respectively. At these depths and for all three boundary tides, the Catskill Shelf, Hudsons Bay Region, and the straits north (Devonian Coordinates) of the Ozark Dome Island are macrotidal whereas the Mid-Continent Shelf is microtidal.

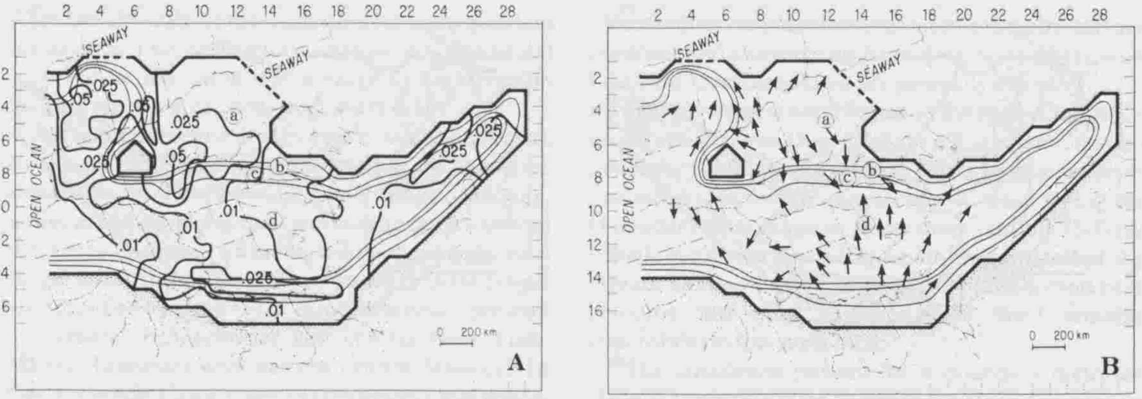


Fig. 7. (A) Residual tidal current magnitudes (m s^{-1}) and (B) directions computed for a boundary tidal range of 3 m and the shelf bathymetry of Fig. 6. Flows are strongest on the edges of the Catskill and Mid-Continent Shelves and directed away from the shelf centres.

Continent Shelf experienced a different tidal regime whose exact nature depends on the boundary tidal range and basin bathymetry.

This dependency is illustrated in Fig. 8 where the average palaeotidal range along Row 16 on the Catskill Shelf for each of 29 computer runs is contoured as a function of boundary tidal range and average thalweg water depth. As discussed previously, the best quantitative estimates of water depth in the Sea are 150–250 m. If the average ocean tidal range today of 1 m is a likely minimum boundary tide, the Catskill Shelf on average is predicted to possess at least a high mesotidal range. Low macrotides are the next most likely. By inference, this applies also to the straits between the Ozark Dome Island and the Transcontinental Arch. It is more difficult to generalize the residual circulation patterns because they change dramatically as a function of assumed bathymetry and palaeogeographic reconstruction.

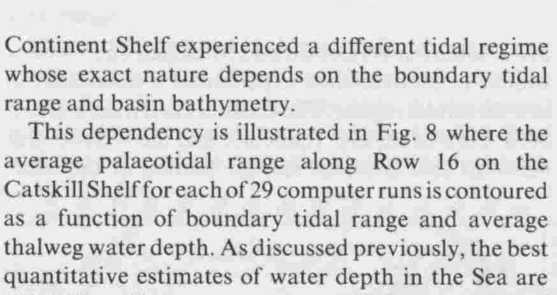


Fig. 8. Summary diagram of the average Catskill Shelf tidal range, as computed by the model, versus the boundary tidal range and average thalweg water depth. For the present best estimates of basin water depths in the range 150–250 m, and boundary tidal range of 1–2 m, the Catskill Shelf experiences high mesotides or low macrotides.

The possible causes for tidal wave augmentation are of two types—convergence, either by landward shoaling or decreasing shelf length, and resonance. Convergence effects can be estimated by the relationship between tidal amplitude, a, alongshore shelf length, l, and shelf water depth, h, for a tidal wave which propagates landward across a frictionless non-reflecting shelf. Because the total energy per unit surface area for a progressive wave is

$$E_{\text{tot}} = \tfrac{1}{2}\gamma a^2, \qquad (5)$$

where γ is the specific weight of water, the total energy for the entire surface area of the wave at any distance, x, from the shelf edge is

$$l_x L_x E_{\text{tot}} = l_x L_x \tfrac{1}{2}\gamma_x{}^2, \qquad (6)$$

where L_x is the tidal wavelength equal to $T\sqrt{gh_x}$, and T is the wave period. Assuming E_{tot} and T are constant as the wave travels shoreward, then

$$l_0 T\sqrt{gh_0}\,\tfrac{1}{2}\gamma a_0{}^2 = l_x T\sqrt{gh_x}\,\tfrac{1}{2}\gamma a_x{}^2$$

where $_0$ denotes a value at the shelf edge, or,

$$\frac{a_x}{a_0} = \left(\frac{l_0}{l_x}\right)^{1/2}\left(\frac{h_0}{h_x}\right)^{1/4}, \qquad (7)$$

a relationship known as 'Green's Law'. Thus tidal wave augmentation should increase with the square root of decreasing shelf length and the quarter root of decreasing shelf depth. For the Catskill Shelf adjacent to a 150 m deep offshore basin, $l_0/l_x \simeq 1{\cdot}5$ and $h_0/h_x \simeq 3$, or $a_x \simeq 1{\cdot}6\,a_0$; that is, much of the augmentation in the model may be explained by convergence effects.

It may be argued that if augmentation is due to convergence, other shelf geometries also fitting the palaeogeographic data may not cause convergence. To consider this possibility, additional solutions were obtained for different clinoform slopes and shelf lengths. Replacing the shelf geometry used in the bulk of the experiments with a uniform $2{\cdot}9 \times 10^{-4}$ slope from the basin edge to a horizontal thalweg floor of 150 m, results in a maximum 8% reduction of the average palaeotidal range along Row 16 on the Catskill Shelf. This insensitivity to shelf bathymetry probably reflects the quarter-root dependency of augmentation on depth. By contrast, modifying the shape of the Catskill shoreline by eliminating the concavity of the embayment along the Catskill Shelf, results in a maximum 43% reduction of the average palaeotidal range along Row 16.

Augmentation due to resonance occurs when the natural period of oscillation on the shelf is coincident with the tidal period. No simple resonance analysis is possible for the Catskill Shelf because the tidal wave in the Catskill Sea propagates parallel to the shelf. Sufficiently deep shelves with tidal waves approaching perpendicularly experience maximum augmentation for across-shelf distances of $\frac{1}{4}, \frac{3}{4}, \frac{5}{4} \ldots$ times L_0. Insofar as this relationship is applicable here, and given that the Catskill Shelf width in the model is approximately $\frac{1}{3}L_0$, then resonance augmentation yields $a_x \simeq 2a_0$ (Slingerland, 1986). Thus resonance effects can also account for most of the augmentation on the Catskill Shelf. Increasing the width while holding depth constant would first decrease the resonance augmentation to negligible amounts and then increase it again as the shelf width approached $\frac{3}{4}L_0$.

Finally, it might be argued that the width of the entrance to the Catskill Sea is not well known and may have been narrower, thus reducing the interior tides. Simulations with the entrance 33% narrower hindcast a 25% lower average palaeotidal range along Row 16 of the Catskill Shelf.

In summary, a hydrodynamic model of the Catskill Sea incorporating the present best estimates of its palaeogeography, palaeobathymetry and palaeogeophysics, indicates high meso- to low macro-palaeotides on the Catskill Shelf along the present Middle Atlantic States. The augmentation with respect to the ocean tide entering the sea occurred due to a combination of convergence and resonance effects and would have been reduced if the Catskill shoreline were less concave, or the shelf were moderately wider, or the Gulf Coast entrance were narrower. These factors in turn would have been controlled by eustatic sea-level or orogenic activity in the rising Acadian Mountains. Thus, one might expect an association between tidal influences and transgressions or regressions, an idea already suggested by McCave (1968) and John S. Bridge (pers. comm.). For example, during transgression the shelf might be more concave, thereby enhancing convergence augmentation. The effects of transgression on resonance augmentation are more complicated—both shelf width and depth would have increased—and remain to be worked out. These results suggest that some facies of the westward prograding Catskill shoreline along the ancestral Appalachian Mountains may owe much of their unique character to tidal influences. Although many facies have been interpreted as wave- or storm-influenced (Sutton, Bowen & McAlester, 1970; Goldring & Bridges, 1973; Goldring & Langenstrassen, 1979), some researchers (Walker & Harms, 1975; Allen & Friend, 1968; Rahmanian, 1979, 1980; Donaldson *et al.* 1984) have observed a lack of well-developed, river-dominated delta facies. Instead they reported extensive mudflat facies (Irish Valley Member) and offshore, shore-parallel sand bar facies (Upper Devonian, First Bradford Formation; Murin & Donahue, 1984) as would be expected in a high mesotidal or low macrotidal setting (Hayes, 1975). Units such as the Luthers Mills Coquinite probably represent tidal channel deposits. Possibly these were deposited in settings and ages of enhanced tides on an otherwise wave (storm)-dominated shelf.

ACKNOWLEDGMENTS

These ideas were nurtured by E. G. Williams, R. J. Cuffey and E. E. Wecker and the manuscript was typed by Judy Bailey. I am indebted to G. deV. Klein and D. L. Woodrow for constructively criticizing the manuscript. Computer funds were provided by the College of Earth and Mineral Sciences, The Pennsylvania State University and the Breen Foundation, Princeton, New Jersey.

REFERENCES

ALLEN, J.R.L. & FRIEND, P.F. (1968) Deposition of the Catskill facies, Appalachian region: with notes on some other Old Red Sandstone Basins. In: *Late Paleozoic and Mesozoic Continental Sedimentation, Northeastern North America* (Ed. by G. deV. Klein). *Spec. Pap. geol. Soc. Am. 106*, 21–74.

BOUCOT, A.J. & GRAY, J. (1983) A Paleozoic Pangea. *Science*, **222**, 571–581.

BRIDGE, J.S. & DROSER, M.L. (1985) Unusual marginal-marine lithofacies from the Upper Devonian 'Catskill clastic wedge'. In: *The Catskill Delta* (Ed. by D. L. Woodrow and W. D. Sevon). *Spec. Pap. geol. Soc. Am. 201*, 143–161.

BRIDGES, P.H. (1982) Ancient offshore tidal deposits. In: *Offshore Tidal Sands* (Ed. by A. H. Stride), pp. 172–192. Chapman & Hall, London.

DENNISON, J.M. (1985) Catskill delta shallow marine strata. In: *The Catskill Delta* (Ed. by D. L. Woodrow and W. D. Sevon). *Spec. Pap. geol. Soc. Am.* 201, 91–106.

DONALDSON, A.C., LEWIS, J.S., MUMCUOGLU, C., BOSWELL, R., PEACE, K. & JEWELL, G. (1984) Upper Devonian depositional systems of Catskill Delta, West Virginia (abstract). *Bull. Am. Ass. Petrol. Geol.* **12**, 1919.

DOTT, R.H., Jr (1974) Cambrian tropical storm waves in Wisconsin. *Geology*, **2**, 243–246.

DRONKERS, J.J. (1964) *Tidal Computations*. North Holland, Amsterdam, 310 pp.

ETTENSOHN, F.R. & BARRON, L.S. (1981) Depositional model for the Devonian–Mississippian Black Shale Sequence of North America: a tectonic-climatic approach. *Morgantown Energy Technology Center, Morgantown, West Virginia, Document 12040–2*, 80 pp.

GOLDRING, R. & BRIDGES, P. (1973) Sublittoral sheet sandstones. *J. sedim. Petrol.* **43**, 736–747.

GOLDRING, R. & LANGENSTRASSEN, F. (1979) Open shelf and near-shore clastic facies in the Devonian. In: *The Devonian System* (Ed. by M. R. House, C. R. Scrutton and M. G. Bassett). *Spec. Pap. Palaeontol.* **23**, 81–97. *Palaeontological Association, London.*

HANSEN, K.S. (1982) Secular effects of oceanic tidal dissipation on the Moon's orbit and the Earth's rotation. *Rev. Geophys. Space Phys.* **20**, 457–480.

HAYES, M.O. (1975) Morphology of sand accumulations in estuaries. In: *Estuarine Research 2* (Ed. by L. D. Cronin), pp. 3–22. Academic Press, New York.

HAYES, M.O. (1979) Barrier island morphology as a function of tidal and wave regime. In: *Barrier Islands* (Ed. by S. Leatherman), pp. 1–27. Academic Press, New York.

HECKEL, P.H. & WITZKE, B.J. (1979) Devonian world paleogeography determined from distribution of carbonates and related lithic paleoclimatic indicators. In: *The Devonian System* (Ed. by M. R. House, C. T. Scrutton & M. G. Bassett). *Spec. Pap. Palaeontol.* **23**, 99–123. Palaeontological Association, London.

HESS, K.W. & WHITE, S.M. (1974) A numerical tidal model of Narragansett Bay. *Univ. Rhode Island Mar. Techn. Rep. No. 20*, 141 pp.

KLEIN, G. DEV. (1974) Estimating water depths from analysis of barrier island and deltaic sedimentary sequences. *Geology*, **2**, 409–412.

KLEIN, G. DEV. (1977) Tidal circulation model for deposition of clastic sediment in epeiric and mioclinal shelf seas. *Sedim. Geol.* **18**, 1–12.

KLEIN, G. DEV. & RYER, P.A. (1978) Tidal circulation patterns in Precambrian, Paleozoic and Cretaceous epeiric and mioclinal shelf seas. *Bull. geol. Soc. Am.* **89**, 1050–1058.

KRAJEWSKI, S.A. & CUFFEY, R.J. (1976) The Ashcraft Limestone Mound. *Pennsylvania Geol.* **7**(5), 9–12.

KRAJEWSKI, S.A. & WILLIAMS, E.G. (1971) Upper Devonian flagstones from northeastern Pennsylvania. *The College of Earth and Mineral Sciences. Spec. Publs Pennsylvania St. Univ. 3–71*, 185 pp.

KROHN, J. & SÜNDERMANN, J. (1982) Paleotides before the Permian. In: *Tidal Friction and the Earth's Rotation II* (Ed. by P. Borsche & J. Sündermann), pp. 190–209. Springer-Verlag, Berlin.

LAMBECK, K. (1980) *The Earth's Variable Rotation: Geophysical Causes and Consequences*. Cambridge University Press, 449 pp.

LEENDERTSE, J.J. (1967) Aspects of a computational model for long-period water wave propagation. *Memo. RM 5294-PR*, Rand Corporation, Santa Monica, CA.

LUNDEGARD, P.D., SAMUELS, N.D. & PRYOR, W.A. (1980) Sedimentology, petrology and gas potential of the Brallier Formation–Upper Devonian turbidite facies of the central and southern Appalachians. *U.S. Dept Energy, Morgantown Energy Tech. Center DOE/METC/5201–5*, 220 pp.

MASCH, F.D. & BRANDES, R.J. (1971) Tidal hydrodynamic simulation in shallow estuaries. *Univ. Texas at Austin, Hydraul. Engin. Lab. Tech. Rep. HYD12–7102.*

MAZZULLO, S.J. & FRIEDMAN, G.M. (1975) Conceptual model of tidally influenced deposition on margin of epeiric seas: Lower Ordovician (Canadian) of eastern New York and southwestern Vermont. *Bull. Am. Ass. Petrol. Geol.* **59**, 2123–2141.

MCCAVE, I.N. (1968) Shallow and marginal marine sediments associated with the Catskill complex in the Middle Devonian of New York. In: *Late Paleozoic and Mesozoic Continental Sedimentation, Northeastern North America* (Ed. by G. deV. Klein). *Spec. Pap. geol. Soc. Am. 106*, 75–107.

MURIN T.M. & DONAHUE, J. (1984) Upper Devonian First Bradford Formation of southwestern Pennsylvania: environment of deposition and factors affecting gas production (abstract). *Bull. Am. Ass. Petrol. Geol.* **68**, 1925.

POTTER, P.E., MAYNARD, J.B. & PRYOR, W.A. (1980) *Final report of special geological, geochemical, and petrological studies of the Devonian shales in the Appalachian Basin.* University of Cincinnati, Ohio, 86 pp.

RAHMANIAN, V.D. (1979) *Stratigraphy and sedimentology of the Upper Devonian Catskill and uppermost Trimmers Rock Formations in central Pennsylvania.* Ph.D. Thesis, The Pennsylvania State University, 340 pp.

RAHMANIAN, V.D. (1980) Upper Devonian tide-dominated deltaic-intradeltaic sedimentation in west-central Pennsylvania: a sedimentologic model for distribution of petroleum sandstone reservoir types (abstract). *Bull. Am. Ass. Petrol. Geol.* **64**, 1288–1289.

ROUSE, H. (1961) *Fluid Mechanics for Hydraulic Engineers.* Dover, New York, 422 pp.

SCHOPF, T.J.M. (1980) *Paleo-oceanography*. Harvard University Press, Cambridge, 341 pp.

SHAW, D.P. (1964) *Time in Stratigraphy*. McGraw Hill, New York, 365 pp.

SLATER, R.D. (1985) A numerical model of tides in the Cretaceous Seaway of North America. *J. Geol.* **93**, 333–345.

SLINGERLAND, R. (1985) Numerical simulation of M_2 co-oscillating tides on very wide mioclinal shelves. *Symp. Modern and Ancient Tidal Deposits*, Utrecht, Netherlands, 26–28 August 1985, *Abs. vol.* 120–122.

SMITH, A.G., HURLEY, A.M. & BRIDEN, J.C. (1981) *Phanerozoic Paleocontinental World Maps*. Cambridge University Press.

SÜNDERMANN, J. & BROSCHE, P. (1978) Numerical computation of tidal friction for present and ancient oceans. In: *Tidal Friction and the Earth's Rotation II* (Ed. by P. Brosche & J. Sündermann), pp. 125–144. Springer-Verlag, Berlin.

SUTTON, R.G., BOWEN, Z.P. & MCALESTER, A.L. (1970) Marine shelf-environments of the Upper Devonian Sonyea Group of New York. *Bull. geol. Soc. Am.* **81**, 2975–2992.

WALKER, R.G. & HARMS, J.C. (1975) Shorelines of weak tidal activity: Upper Devonian Catskill Formation, Central Pennsylvania. In: *Tidal Deposits* (Ed. by R. N. Ginsburg), pp. 103–108. Springer-Verlag, New York.

WEBB, D.J. (1982) On the reduction in tidal dissipation produced by increases in the Earth's rotation rate and its effect on the long-term history of the Moon's orbit. In: *Tidal Friction and the Earth's Rotation II* (Ed. by P. Brosche & J. Sündermann), pp. 210–221. Springer-Verlag, Berlin.

WOODROW, D.L., ENOS, P., BOTTJER, R.J. & MINERO, C. (1981) Bioclastic carbonate units in the Catskill clastic wedge. In: *Guidebook for Field Trips in Southcentral New York* (Ed. by P. Enos), pp. 201–225. New York State Geological Association, 53rd Annual Meeting.

WOODROW, D.L., FLETCHER, F.W. & AHRNSBRAK, W.F. (1973) Paleogeography and paleoclimate at the depositional sites of the Devonian Catskill and Old Red Sandstone Facies. *Bull. geol. Soc. Am.* **84**, 3051–3064.

WOODROW, D.L. & ISLEY, A.M. (1983) Facies, topography, and sedimentary processes in the Catskill Sea (Devonian), New York and Pennsylvania. *Bull. geol. Soc. Am.* **94**, 459–470.

ZIEGLER, A.M., SCOTESE, C.R., MCKERROW, W.S., JOHNSON, M.E. & BAMBACH, R.K. (1979) Paleozoic paleogeography. *Ann. Rev. Earth planet. Sci.* **7**, 473–502.

(Manuscript received 21 February 1985; revision received 10 July 1985)