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Thermal Contraction and Flexure of Mid–Continent and Atlantic Marginal Basins

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Summary

Thermal contraction of the lithosphere is a probable cause of the gradual subsidence indicated by sediments of mid-continent basins and Atlantic continental shelves. The subsidence is complicated by time dependent regional isostatic compensation since adjacent parts of the lithosphere are mechanically coupled, and since creep in the lithosphere may relieve accumulated stress. Thus, if more subsidence occurs at point A than nearby point B, point A would be buoyed up and point B dragged down. Relaxation of this coupling during a later period of more gradual subsidence would produce uplift at B and downwarp at A. The absence of younger beds over local minima of subsidence such as the Florida arch, the flanks of the Michigan basin, and the Atlantic coastal plain (USA) can thus be explained. Variations in the subsidence rate due to exponential decay of the thermal anomaly or to starved basin-evaporite depositional sequences can produce observable effects.

Analytic models of the Michigan basin and the Atlantic coast (USA) are compatible with previously estimated parameters: thermal decay time of the lithosphere, 50 My; flexural parameter of the lithosphere beneath air, 200 km; and viscosity of the lithosphere, 10^{25} poise. The effects of flexure are not clearly evident in Silurian evaporite deposition in the Michigan basin and it is probable that an extended time was required for the evaporite sequence to accumulate.

The cause of the thermal heating event which precedes subsidence is unclear for mid-continent basins although bulk replacement of the uppermost mantle is necessary. The heating events may be associated with periods of slow sea-floor spreading (when slabs exert a tensional force on the lithosphere) and hence low eustatic sea level. There is little direct evidence that an initial heating event actually occurred in the Michigan basin immediately before the start of subsidence.

Introduction

A few kilometres of gently dipping shallow water sediments in the interior of stable continental regions (such as the Michigan basin) and along stable Atlantic

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continental margins indicate vertical tectonic movements during the period of sediment accumulation. Purely eustatic sea-level changes cannot be the primary cause of sedimentation, as the thicknesses of sediment vary laterally from zero to a few kilometres. The areas of greatest sediment preservation are not synclines formed by folds in originally horizontal strata. The presence of deeper water sediments and thicker units near the centres of continental basins indicates that the basin was an area of rapid subsidence during deposition. The similarity of geometry and history of differential subsidence of Atlantic margins and mid-continent basins has been noted for some time (for example, Kay 1951). Not only are the thicknesses of beds similar, but the distance from flank to centre of mid-continent basin is about equal to the width of continental shelf basins (200–300 km). Also, the subsidence rate in both regions can be approximated by an exponential function which decreases with a time scale of about 50 My (Sleep 1971). This gross similarity between continental shelf and intracratonal basins occasions use of similar working hypotheses for the mechanics of subsidence in both regions.

Theories explaining the subsidence of Atlantic continental margins are based on the origin of these margins when a pre-existing continent was broken up by the initial spreading of a new mid-ocean ridge. Immediately after breakup, rapid subsidence at the continental margin may occur due to thinning or loading of the continental crust either by necking, loading by intrusions or subcrustal erosion. A different mechanism is needed to explain the slow continual subsidence indicated by the shelf and mid-continent basin deposits.

In the paper we mathematically model the hypothesis that gradual thermal contraction of the lithosphere causes the subsidence (Vogt & Ostenso 1967; Schneider 1969, 1972; Sleep 1971, 1973; Keen & Keen 1973), explicitly including regional

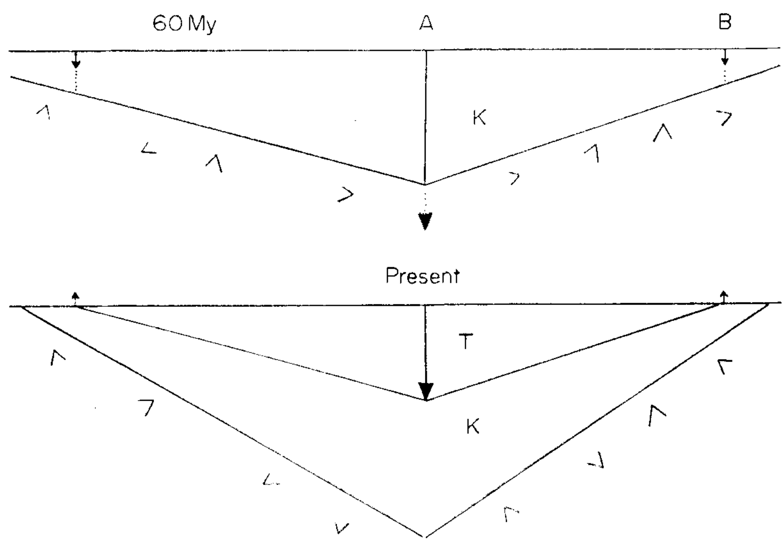

FIG. 1. Schematic diagrams show effect of regional isostasy on subsidence. During the Cretaceous period (above) more subsidence occurred at point A than point B. Point A is buoyed up and point B dragged down by the amount indicated by dashed lines. During the Tertiary period less subsidence occurred. The regional isostatic stresses built up during the Cretaceous relax causing uplift at point B and additional subsidence at point A. Note the Cretaceous beds outcrop at point B (Modified after Sleep 1973).

isostatic effects.* The qualitative reason for including this coupling of adjacent points can be seen as follows (McGinnis 1970; Walcott 1972a; Sleep 1971, 1973). If more subsidence during an earlier period (say the Cretaceous) occurred at point A than at nearby point B, point A is buoyed up and point B dragged down by the amount indicated by dotted lines (Fig. 1). During the Tertiary period the rate of loading of the lithosphere by thermal contraction would be less, because cooling of a plate obeys a decreasing exponential law. If the lithosphere is subject to creep, the regional isostatic stress built up during the Cretaceous would relax causing uplift at point B and additional subsidence at point A. The mechanism provides a possible explanation for the observed outcrop of older beds near minima of subsidence. In addition to thermal contraction, any process that is localized in space and episodic in time can cause this effect. Information on the rheological properties of the lithosphere can be obtained by comparing mathematical models of this process with observed data.

We construct mathematical models explicitly resembling the Michigan basin in post-Sauk (Middle Ordovician and later times) and incidentally applicable to the east coast of the United States in Cretaceous and Tertiary times. Although it is harder to explain or to detect the initial heating episode in a closed mid-continent basin, such a region is simpler to analyse than an Atlantic margin. The weight of sediment on the ocean floor may drag down the nearby continental shelf (Dietz 1963; Walcott 1972a), or the continental crust may flow out over the more dense oceanic lithosphere (Bott 1971, 1973; Bott & Dean 1972). We do not attempt to analyse the possibility that migration of interstitial melt to and from the asthenosphere causes subsidence (Scheidegger & O'Keefe 1967; Sloss & Speed 1974).

Subsidence of filled basin

Following Walcott (1970a, b, c) we model the elastic behaviour of the lithosphere as a viscoelastic slab. This mathematical treatment was developed for engineering purposes by Nadai (1961). In one horizontal dimension the displacement of a loaded viscoelastic slab is governed by a fourth order partial differential equation (Nadai 1961, p. 285),

$$N\frac{\partial^4}{\partial x^4}\left(\frac{\partial w}{\partial t}\right) = \left(\frac{\partial}{\partial t}+\frac{1}{t_e}\right)p_0 \tag{1}$$

where w = vertical displacement, positive downward,

x = horizontal co-ordinate,

t = time, future positive,

p_0 = load per area,

N = elastic flexural rigidity, and

t_e = viscoelastic decay time, positive.

In terms of the material properties

$$t_e = 3\eta/E \tag{2}$$

and

$$N = Eh^3/9 \tag{3}$$

* No attempt is made here to review the voluminous older literature on this subject. The relevant equations have been known generally for over 70 years and the sign of the effect is obvious. The contribution herein is to apply the equations to a situation made patent by plate tectonics. See Gunn (1949) for earlier references and a discussion of non-time dependent regional isostasy.

where η = shear viscosity of slab,

E = Young's modulus of slab, about 10^{12} dyne cm^{-2}, and

h = thickness of slab.

For an isostatically compensated lithosphere, the asthenosphere causes an additional load proportional to the deflection of the slab and the specific weight contrast between the asthenosphere and the material (be it water or sediments) filling the depression.

Here,

$$p_0 = p - \Delta\rho g w \tag{4}$$

where p is load that is independent of displacement and

$\Delta\rho g$ is the contrast in specific weight between asthenosphere and sediments or water, positive.

Letting $\Delta\rho g = k$ equation (1) becomes

$$N \frac{\partial^4}{\partial x^4}\left(\frac{\partial w}{\partial t}\right) = -k\left(\frac{\partial w}{\partial t} + \frac{w}{t_e}\right) + \frac{\partial p}{\partial t} + \frac{p}{t_e}. \tag{5}$$

Following Nadai (1961, p. 301), we solve (5) by assuming a sinuoidal spatial variation for w and combining such solutions using Fourier series.

We set

$$\left.\begin{aligned} w &= \phi_m(t)\cos\left(\frac{m\pi m}{b}\right) \\ &\text{and} \\ p &= p_m \cos\left(\frac{m\pi x}{b}\right)\Psi_m(t), \end{aligned}\right\} \tag{6}$$

where ϕ_m, Ψ_m = functions of time,

m = positive integer, and

b = positive length.

A cosine series is used since our model basin is assumed symmetric about $x = 0$. Substituting (6) into (5) we obtain an equation for ϕ;

$$N \frac{m^4\pi^4}{b^4}\frac{\partial \phi_m}{\partial t} = -k\left(\frac{\partial \phi_m}{\partial t} + \frac{\phi_m}{t_e}\right) + p_m\left(\frac{\partial \Psi_m}{\partial t} + \frac{\Psi_m}{t_e}\right). \tag{7}$$

Letting $t_m \equiv \left(1 + \dfrac{\pi^4 m^4 N}{kb^4}\right) t_e$, equation (7) can be written as (8)

$$t_m \frac{\partial \phi_m}{\partial t} + \phi_m = C_m\left(\frac{\partial \Psi_m}{\partial t} t_e + \Psi_m\right), \tag{9}$$

where $C_m = p_m/k$.

This equation (9) can be solved to obtain (Nadai 1961, p. 306)

$$\phi_m = D_m e^{-t/t_m} + \frac{C_m}{t_m}\left[t_e \Psi_m + \left(1 - \frac{t_e}{t_m}\right) e^{-t/t_m}\int_0^t \Psi_m e^{s/t_m} ds\right]. \tag{10}$$

The quantity t_m is thus the decay time for an initial deflection of wavelength $2b/m$. From equation (8) it can be seen that the decay time for a long wavelength deflection is approximately t_e and the decay time for a short wavelength deflection is approximately proportional to the inverse fourth power of the wavelength.

We define a length scale, called the *flexural parameter* by Walcott (1970a, b, c),

$$\alpha = (4\,N/k)^{\frac{1}{4}}. \tag{11}$$

In a region away from a concentrated load, the deflection of an elastic slab is solved by Nadai (1961, p. 288).

$$w = \exp\,(-x/\alpha)\cos\,(x/\alpha). \tag{12}$$

The parameter α is thus directly determined from studies of deflections near the termini of glaciers and found to be about 200 km in central North America (Walcott 1970a, b, c). For a basin filled with sediments, the parameter α would be increased to 300 km since $k = 0{\cdot}6$ to $0{\cdot}7 \times 10^3$ dynes cm^{-3} rather than $3{\cdot}3 \times 10^3$ dynes cm^{-3} for the depressions near glaciers, which for the most part remain unfilled.

Load due to thermal contraction

Considering the lithosphere as a thin plate causes some mathematical difficulty in modelling loads due to thermal contraction because the deflection, w_c, at the middle of the plate should be included on the left side of equations (1) and (5), while the deflection of the free surface, w_s, controls loads due to sediments on the right side of equations (4) and (5). We obtain an expression, correct to the first order in the excess temperature, by summing loads and deflections in a co-ordinate system centred on the middle of the plate, depth $= h/2$. Assuming thermal contraction is isotropic, horizontal contraction will increase the weight per area of a column through the lithosphere by

$$p_{\mathrm{H}} = -\tfrac{2}{3}\gamma\rho\, gh T_{\mathrm{E}}, \tag{13}$$

where T_{E} = average excess temperature, and γ = volume coefficient of thermal expansion, positive constant. If vertical thermal contraction is equally distributed above and below the middle of the plate, the deflection at the base of the plate will be given by

$$w_{\mathrm{b}} = w_{\mathrm{c}} + \tfrac{1}{6}\gamma h T_{\mathrm{E}} \tag{14a}$$

and the deflection at the free surface is given by

$$w_{\mathrm{s}} = w_{\mathrm{c}} - \tfrac{1}{6}\gamma h T_{\mathrm{E}}. \tag{14b}$$

The load due to deflection of the base of the plate is

$$p_{\mathrm{b}} = -\rho g w_{\mathrm{b}}; \tag{15a}$$

and the load due to sediments at the free surface is

$$p_{\mathrm{s}} = \rho_{\mathrm{s}}\, g w_{\mathrm{s}} = (\rho g - \Delta\rho g)\, w_{\mathrm{s}}, \tag{15b}$$

where ρ_{s} is the density of sediments.

By adding p_{H} in equation (13) and p_{b} and p_{s} in equation (15) and applying equation (14), we obtain an expression similar to equation (4), given by

$$p_0 = -\gamma T_{\mathrm{E}}\, h(\rho g - \tfrac{1}{6}\Delta\rho g) - \Delta\rho g w_{\mathrm{c}}. \tag{16}$$

In the case of pointwise isostacy, p_0 is zero and the elevation, as expected, is given by

$$w_{\mathrm{s}} = -\frac{\rho g}{\Delta\rho g}\gamma T_{\mathrm{E}}\, h. \tag{17}$$

From (14b) it can be seen that 1/6 of the thermal contraction of the plate produces a direct increment to w_{s} rather than a load in equation (16). The fractional amount of subsidence thereby produced is minor, especially if $\Delta\rho g$ is small. To permit rescaling and to simplify the equations, we will let w represent w_s and the load,

$$p = -\gamma T_E h\rho g, \tag{18}$$

throughout the remainder of the paper. This approximation corresponds physically to assuming that resistance to flexure is concentrated near the surface. We also ignore all second order temperature changes from vertical displacement of the plate and accumulation of sediment.

Solution of equations

To obtain the load function Ψ_m in equations (6)–(10), we note that, ignoring higher vertical harmonics, the temperature in a two-dimensional slab having isothermal boundaries can be represented as

$$T_e(x, z) = \exp(-\kappa\pi^2 t/h^2) \sin(\pi z/h) \sum_0^\infty C_m \exp(-\kappa\pi^2 m^2 t/b^2) \cos(m\pi x/b), \tag{19a}$$

where κ = thermal diffusivity, a positive constant, and T_e = excess temperature, relative to final equilibrium temperature.

Because $hT_E \equiv \int_0^h dz T_e$, we find from (18) that

$$\Psi_m(t) = (1 - \exp(t/t_{pm})), \tag{19b}$$

where

$$t_{pm} \equiv [\kappa\pi^2(1/h^2 + m^2/b^2)]^{-1}. \tag{19c}$$

As shown below, horizontal conduction of heat is minor in the case of interest. The dependence of Ψ_m on m can be thus ignored to give,

$$\Psi = (1 - e^{-t/t_p}). \tag{20}$$

Substituting (20) into (10), we obtain an expression for ϕ_m (Nadai 1961, p. 307),

$$\Phi_m \equiv \frac{\phi_m}{C_m} = \frac{1}{t_p - t_m} [(t_p - t_e)(1 - e^{-t/t_p}) - (t_m - t_e)(1 - e^{-t/t_m})]. \tag{21}$$

To obtain expressions for C_m, we will assume that the load in (18) can be represented by a gaussian curve with unit amplitude,

$$\frac{p}{k} = e^{-x^2/a^2}(1 - e^{-t/t_p}), \tag{22}$$

where a is the gaussian width of the eventual basin. The gaussian function gives a good representation of actual basins (Sloss & Scherer 1975) and yields tractable mathematics. The width so defined of the Michigan basin is about 170 km across strike.

From Fourier analysis of (22), we obtain

$$C_m = \frac{2}{b} \int_0^b \cos\left(\frac{m\pi x}{b}\right) e^{-x^2/a^2} dx \tag{23}$$

and

$$w = C_0 + \sum_{m=1}^\infty C_m \Phi_m(t) \cos\left(\frac{m\pi x}{b}\right) \tag{24}$$

The form given is numerically equivalent to the exact solution obtainable by Fourier integral and more convenient to evaluate if $b \gg a$. The upper limit of integration in (23) then can be set to infinity (without affecting the accuracy obtainable on a computer) to obtain a closed form for C_m. In our evaluation, we let $b = 10a$ and set C_0 such that $w(b) = 0$.

The value of the viscous decay time, t_e, determined by Walcott (1970a, b, c) from the apparent flexural parameter for features of various geological ages is about 10^5 to 10^6 yr. The time constant for thermal contraction t_p is about 50 My (Sleep 1971). For the large times (compared with t_e) on which platform deposits accumulate, the term containing t_e on the right side of (9) is small compared to the other term on that side for wavelengths which have not already attenuated ($t_m \gg t_e$). Equation (9) thus becomes

$$\frac{\pi^4 m^4 N}{kb^4} t_e \frac{\partial \phi_m}{\partial t} + \phi_m = C_m \Psi, \tag{25}$$

or, inserting (2) into (25),

$$\frac{\pi^4 m^4 h^3 \eta}{3kb^4} \frac{\partial \phi_m}{\partial t} + \phi_m = C_m \Psi, \tag{26}$$

which is the expression which would have been obtained had elastic effect been ignored initially (Nadai 1961, p. 286). Equation (21) indicates that the viscosity of the lithosphere cannot be deduced from long duration processes alone and that we need the flexural parameter for short durations or some other additional information. As the elastic Young's modulus E and the flexural parameter α are reasonably well determined, we can express the decay time for a persistent ($t_m \gg t_e$) wavelength as

$$t_m = \frac{3\pi^4 \alpha^4 \eta}{4a^4 E} \left(\frac{a^4 m^4}{b^4}\right) = \frac{\pi^4 \alpha^4}{4a^4} \left(\frac{a^4 m^4}{b^4}\right) t_e. \tag{27}$$

The importance of knowing the ratio of the flexural parameter (α) to the width of the basin (a) should be evident from this equation. As the parameters m and b exist only for the Fourier analysis, the physically significant parameter is a/α.

This form of equation (27) is useful for justifying our assumption that lateral heat transport is unimportant, contrary to some speculation (Long & Lowell 1973). Firstly, t_m varies as the fourth power of wavelength in contrast to the thermal loading parameter, t_{pm}, which varies as the square of wavelength for small wavelength. Secondly, t_{pm} begins to vary rapidly only for half-wavelengths less than the thickness of the lithosphere, about 100 km, while t_m varies rapidly for all wavelengths of interest. Finally, thermal anomalies with very short horizontal wavelength are not strongly excited by our choice of basin shape and in nature would have probably cooled before deposition began anyway. Ignoring lateral heat transport permits rescaling described below.

We ignore in the model parameters any possible differences between the thermal properties of the oceanic lithosphere, inferred from subsidence of mid-oceanic ridges, and continental lithosphere, which we wish to model. The bulk of both lithospheres is composed of mantle of probably similar composition and properties. In any case, our results may be easily rescaled in time by multiplying the decay time, t_e, the thermal loading time, t_p, and the time scale of our results by a constant factor, if a different time constant is warranted by the data.

We also ignore thermoelastic stresses with some justification, but largely because we are not certain of the means by which we should include them. To the first order, no bending moment results from the cooling of continental lithosphere, since the primary thermal anomaly is centred vertically such that the middle of the lithosphere cools with respect to top and bottom. Also, thermal elastic stress, unlike the stress due to spatial distribution of load, vanishes once it is relaxed after time t_e.

Use of a model with one horizontal dimension leads to an overly rapid relaxation of flexural stresses and an overestimate of the viscosity of the lithosphere when the model is compared to a real two-dimensional basin. For loads having a large value of t_m, the error in calculated t_m and viscosity cannot be greater than a factor of 4, as can be seen if $w = \cos(x/b)\cos(y/b)\,\phi_m(t)$ is inserted into (5) and $\partial^4/\partial x^4$ is replaced by $(\partial^2/\partial x^2 + \partial^2/\partial y^2)^2$.

The present position of strata for models representing the ages of the Atlantic Coast of the United States and the Michigan Basin were calculated using (23) and (24) and various values of α/a and

loading time constant $t_p = 50$ My,

viscoelastic decay time $t_e = 1$ My, and

initial load $D_m = 0$.

The results are plotted in Figs 2 and 3. It should be noted from (27) that a factor of 2 increase in α/a can be traded off for a factor of 16 increase in viscosity, η, or relaxation time, t_e. Large values of α/a or t_e result in incomplete subsidence at the centre of the basin, which could be detected from gravity anomalies (McGinnis 1970; Walcott 1970a, 1972b). If α/a is too small, the characteristic outcrop of progressive older beds toward the edge of the margin is not produced.

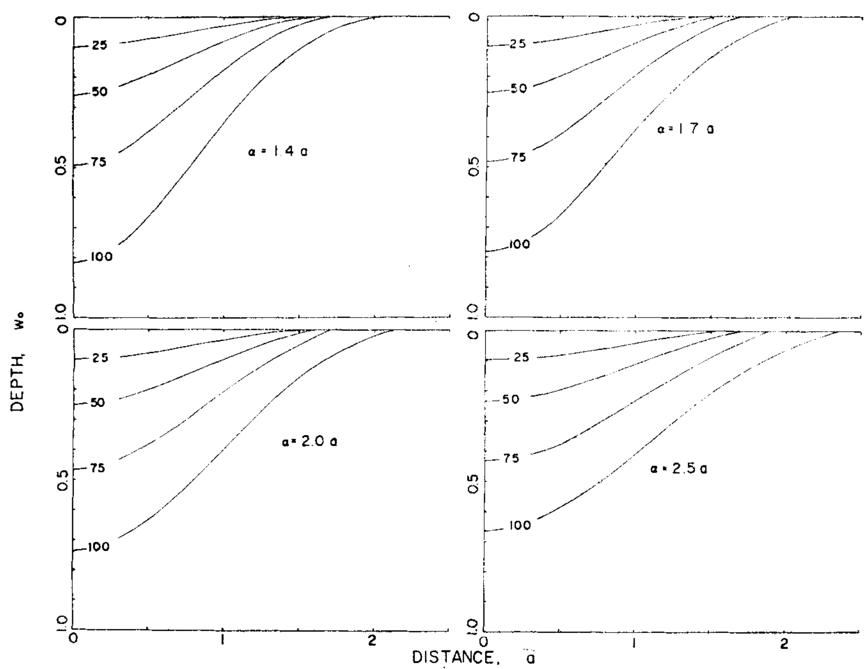

FIG. 2. Theoretical cross-sections of two-dimensional, symmetric basins after 100 My of subsidence are plotted for various ratios of the characteristic basin width (a) and the flexural parameter (α). Numbers indicate age of beds. A viscoelastic decay time of 1 My or equivalently a viscosity of the lithosphere of 10^{25} poise is assumed. The patterns from $\alpha = 1{\cdot}4\ a$ to $\alpha = 2{\cdot}5\ a$ are acceptable for the Atlantic Coast. Flexural upward of the flanks is insignificant for lower values of α (or viscosity). Large values of α lead to an excessive remaining load at the centre of the basin which is not indicated by gravity anomalies. As the observed ratio of α (= 300 km for subsidence beneath sediments) and a (= 150 km for Atlantic margin) is about 2, the results are compatible with the assumed viscosity.

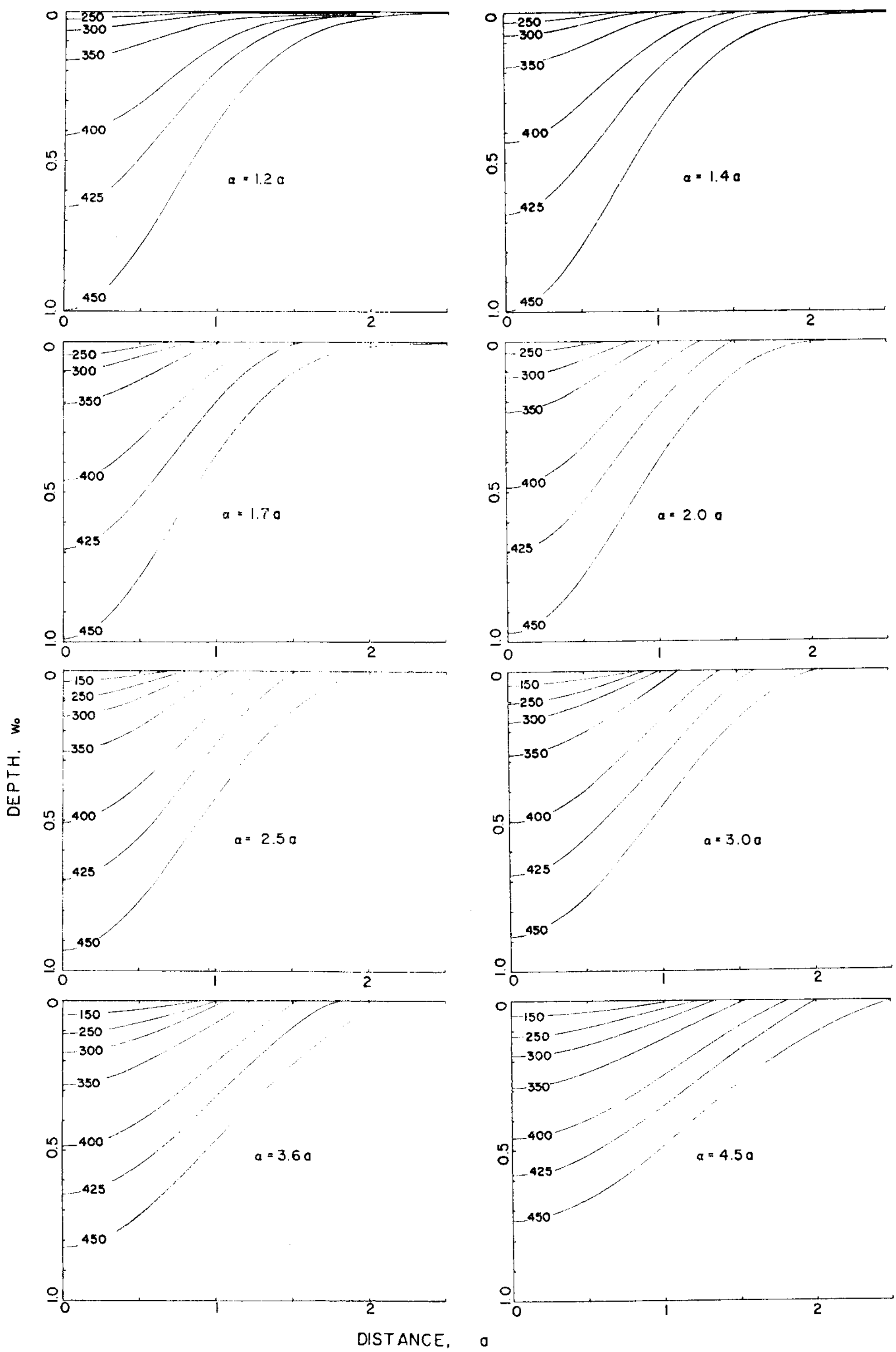

Fig. 3. Theoretical subsidence curves after 450 My are plotted as in Fig. 2. The patterns from $\alpha = 1 \cdot 4\ a$ to $\alpha = 2 \cdot 5\ a$ are acceptable for the Michigan basin. As the observed ratio of α (300 km) and a (= 170 km for the Michigan basin) is about $1 \cdot 7$, results are compatible with the assumed viscosity of 10^{25} poise.

Episodically filled basin

Physical basins do not remain filled to the brim as we have so far assumed. To further test our mechanics of subsidence, we construct a model intended to be relevant to sediment starvation of the Michigan basin during Silurian times and the rapid deposition which subsequently filled the basin.

The qualitative sequence of events resulting from episodic rather than continuous supply of sediments can be understood by considering a basin which is sediment-starved near its centre for a period of time and then is rapidly filled. The bottom of the basin subsides slowly if the interior of the basin does not receive a sediment load. When the basin is quickly filled with sediments, loading is concentrated near the centre. Flexural coupling of the centre to the flanks of the basin causes the centre to be buoyed up and the flanks to be dragged down. As with the previous example of a basin which was continually filled, viscous relaxation of the regional isostatic compensation of the load would cause the flanks to move upward and the centre to subside during subsequent periods of time. The effect in the case of a rapidly filled starved basin would be pronounced, as the subsidence due to loading is more rapid and spread over a larger area. It can, therefore, be expected that transient flexural effects of intermittent sedimentation would be superimposed noticeably on the general flexural effects.

To quantify this mechanism it is necessary to formulate and solve the equations of subsidence in a basin which does not always remain filled with sediments. For purposes of discussion we assume the following subsidence history of a basin having one horizontal dimension.

1. At time $t = 0$

   No vertical displacement, $w = 0$,

   No load, $p = 0$,

   No sediment, $s = 0$,

   No topography, $D = 0$.

The load due to thermal contraction as in (22) is assumed to be

$$\frac{p}{k} = (1-e^{-t/t_p})\, e^{-(x^2/a^2)}. \tag{28}$$

2. Basin remains filled until $t = t_1$ when $w = s = w_1 = s_1$, $p = p_1$

3. For $t_1 \leqslant t \leqslant t_2$,

$$p_0 = [(p_1 - kw_1) + (p - p_1) - k(x)(w - w_1)], \tag{29a}$$

where k is specific weight contrast for filled basin.
This assumes that ratio of accumulation of sediments to subsidence is only a function of position while the basin is starved. Thus,

$$\left(\frac{s-s_1}{w-w_1}\right) = \frac{k(x)-k_m}{k-k_m} \tag{29b}$$

where k_m is the load due to subsidence beneath water about $2{\cdot}3\times 10^3$ dynes cm^{-3}. The water depth is computed by

$$D = (w-w_1)\frac{k-k(x)}{k-k_m}. \tag{30}$$

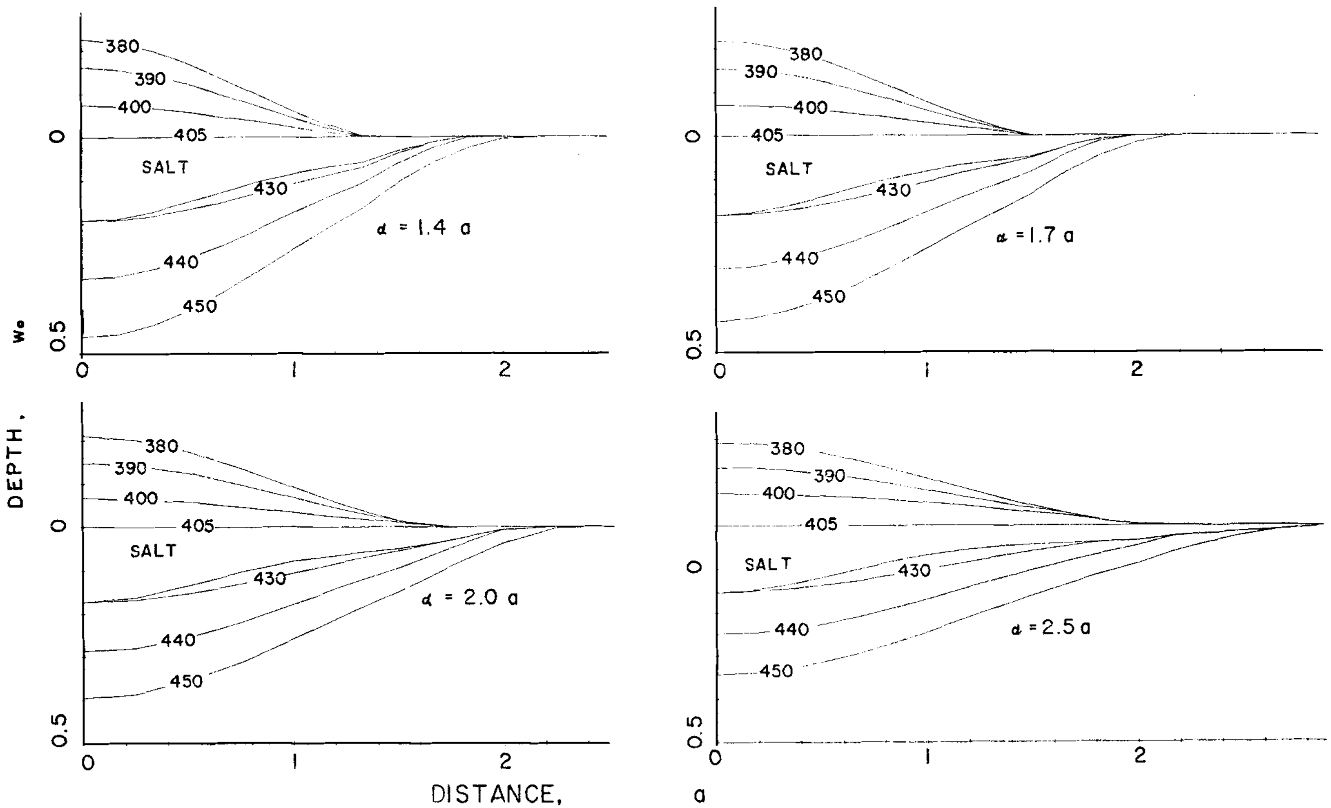

FIG. 4. Theoretical subsidence curves are plotted as in Fig. 2 except that the top of salt which instantaneously filled the basin at 405 My and the erosional surface at 380 My are used as a basinline. From 430 to 405 My the basin was starved with $\beta = 1$. The basin was considered full at other times. Extensive prograding of the salt over the flanks of the basin occurs in the figure but not in the Michigan basin. The unconformity over the salt is also an onlap-offlap rather than the gradual offlap in the figure.

For purposes of calculations we let the fraction of the deflection during the period when the basin is starved be

$$F = 1 - \exp(-x^2/\beta^2). \tag{31}$$

This corresponds to

$$k = (0{\cdot}7 + 1{\cdot}6 \exp(-x^2/\beta^2))\ \mathrm{gm\ cm^{-3}\ g}, \tag{32}$$

$$s = w_1 + (w(t) - w_1)\,F, \quad \text{and} \tag{33}$$

$$D = w - s, \quad \text{where} \tag{34}$$

D = water depth,

s = sediment thickness, and

β = width of starved region.

4. At t_2 the basin became filled either instantly or over a period of time. The instantaneous case can be treated simply by the analytical method described above

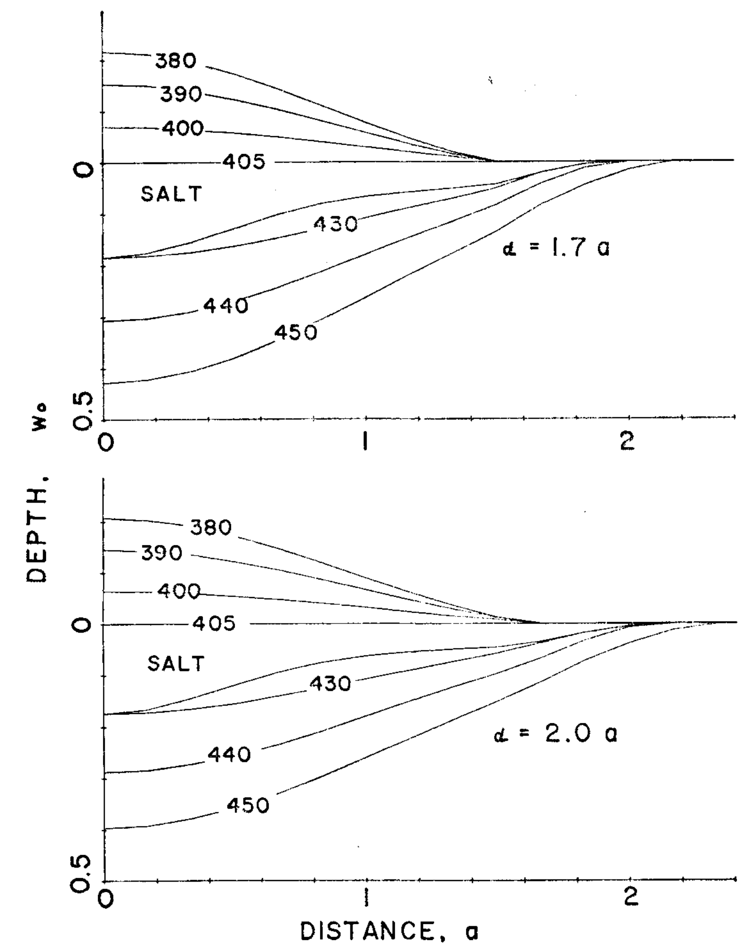

FIG. 5. Theoretical subsidence curves are plotted as in Fig. 4. In this calculation the basin was less starved with $\beta = 1/\sqrt{2}$. The results are similar to the previous figure.

using the values of w and p at time $= t_2$ as initial conditions. Elastic deformation of the lithosphere is caused by the new sediment load changing the basin geometry such that the basin is filled and in mechanical equilibrium. This deflection, w_e, is computed by

$$N\frac{\partial^4}{\partial x^4}(w_e) = -kw_e + 1\cdot 6\,\mathrm{gm\,cm^{-3}}\,D(t_2)\,g. \tag{35}$$

Any physical filling of a basin which is rapid compared to the viscoelastic time constant t_e can be modelled as being instantaneous. Slower filling of the basin can be modelled numerically by calculating the net amount of sediment which would be needed to fill the basin as above, but only adding a fraction of this amount of sediments. This process then can be repeated several times until the basin is nearly full at which time the remainder is filled instantaneously as before. We ignore differences in densities of the sediments and the evaporites which are assumed to fill the starved basin.

Equations (5) and (35) can be easily solved numerically using finite difference methods. This method entails replacing derivatives by differentials. For example,

$$\frac{\partial w}{\partial t} = (w_{u+1} - w_u)/\Delta t, \tag{36}$$

where u (subscript) indicates the time step and elapsed time $-u\Delta t$. This function, w, is thus defined only on a discrete set of grid points in x and time steps in t.

At any time interval we know p by assumption and p and w from previous steps. We wish to find w at the next time step, t_{u+1}. Using (5), (28), and (32), we obtain

$$\frac{N\left(\left(\frac{\partial^4 w}{\partial x^4}\right)_{u+1} - \left(\frac{\partial^4 w}{\partial x^4}\right)_u\right)}{\Delta t} = -k(x)\left(\frac{w_{u+1}-w_u}{\Delta t} + \frac{w_u}{t_e}\right) + \left(\dot{p}(x) + \frac{p(x)}{t_e}\right) \tag{37}$$

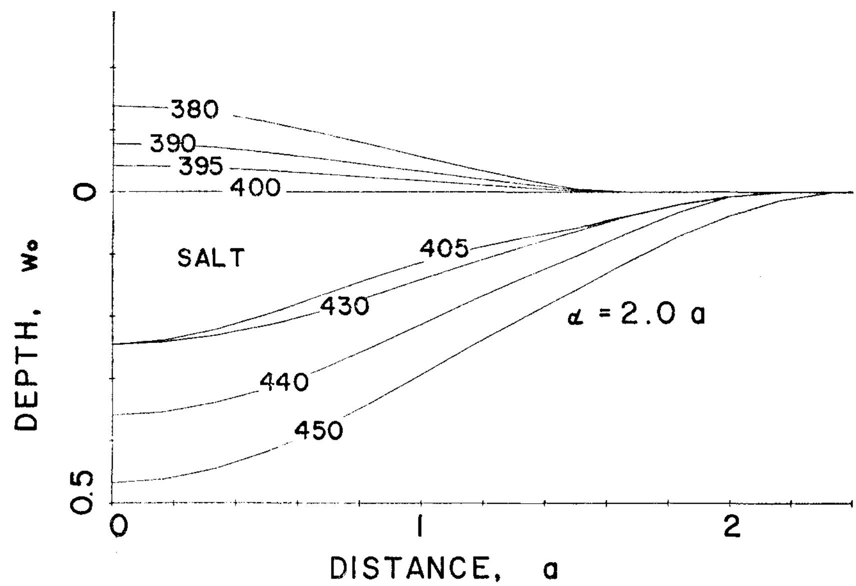

FIG. 6. Theoretical subsidence curves are plotted as in Fig. 4, except that the 400 My time line is used as a baseline. Salt was assumed to gradually fill the basin between 405 and 400 My. Subsequent rebound and subsidence are only slightly different from Fig. 4.

Expanding the derivatives in (37) and letting $u+1$ and u be time steps, and m be horizontal position, $x = m\Delta x$, we obtain

$$\frac{N(6w_{m,\,u+1}+w_{m+2,\,u+1}+w_{m-2,\,u+1}-4w_{m+1,\,u+1}-4w_{m-1,\,u+1})}{(\Delta x)^4\,\Delta t}$$

$$-\frac{N(6w_{m,\,u}+w_{m+2,\,u}+w_{m-2,\,u}-4w_{m+1,\,u}-4w_{m-1,\,u})}{(\Delta x)^4\,\Delta t}$$

$$= -k(x)\left(\frac{w_{u+1}-w_u}{\Delta t}+\frac{w_u}{t_e}\right)+\left(\dot{p}(x)+\frac{p(x)}{t_e}\right). \quad (38)$$

This equation is then solved for $w_{m,\,u+1}$ for each point. The new values of $w_{m,\,u+1}$ are then used to repeat the solution at each point. This process, known as pointwise relaxation, is repeated until it converges. The values of $w_{m,\,u+1}$ are then used to repeat the process at the next time step. The main restriction is that the time step must be small compared to t_e and $p/\dot{p}$ to insure stability.

For purposes of computation we let the grid be

(a) $\Delta x = 0{\cdot}25\,a$

$b = 7{\cdot}5\,a$

or

(b) $\Delta x = 0{\cdot}167\,a$

$b = 5\,a.$

Boundary conditions were

$$\text{at } x = 0,\ \partial w/\partial x = \partial^3 w/\partial x^3 = 0,$$

and

$$\text{at } x = b,\ w = \partial w/\partial x = 0.$$

The numerical scheme was checked against the analogous analytic solution for filled basins. Runs using grids (a) or (b) were checked against each other. Grid (b) gave slightly better results although both numerical and analytic schemes agree to better than 2 per cent for small times and 1 per cent after about a time of $15t_e$.

The results of the calculations are plotted in Figs 4, 5, and 6. As with the case of a filled basin, flexural effects become more pronounced if the ratio flexural parameter to basin width is high. While the basin is starved, the flanks of the basin rebound as stress accumulated during earlier stages of subsidence is relaxed. The load due to rapid accumulation of salt results as expected in a broad downwarp and subsequent rebound of the flanks. The results are not sensitive to the extent to which the basin is starved (Figs 4 and 5), nor to whether the accumulation of the salt is gradual over 5 My rather than instantaneous (Fig. 6). Calculations not shown indicated that the rescaling with respect to α/a and η can still be done except immediately after salt deposition.

It should also be noted that the geometry of strata returns after several million years to that of a continually filled basin. This gives added justification in using filled basin models for a first-order approximation of the history of basins having episodic sedimentation.

Comparison with observations

We test and calibrate our mechanical models by comparing them with data on the Cretaceous to Recent subsidence of the east coast of the United States and the Middle Ordovician to Jurassic subsidence of the Michigan basin.

For the most part we are concerned with sedimentary units thick enough to average out variations in sediment supply and eustatic variations of sea level. Beds deposited during eustatic transgressions are subject to erosion during later regressions and are more completely preserved in regions of rapid subsidence. The bowl shape of sequences defined by major transgressive events is thus enhanced by erosion, as well as by syn-depositional flexure (See Sloss 1972). We can infer from this that the Michigan Basin and the Atlantic shelf basin generally remained nearly filled with sediments since the eustatic changes produce unconformities, while an unfilled basin would receive an increased supply of sediments at times of low sea level. Quantification of lithofacies variations, the amplitude of the eustatic curve, and determination of whether eustatic variations are the result of processes beneath the land or sea are beyond the scope of this paper.

The amount of excess temperature needed to produce the observed subsidence of 3–4 km in the Atlantic marginal basins and the Michigan basin can be inferred by comparison with the subsidence at mid-oceanic ridges where the average excess temperature is about 600 °C and the subsidence is about 3 km. Using equation (17) and noting that $\Delta\rho g$ is about $2{\cdot}3\times10^3$ dynes cm^{-3} for ridges and about $0{\cdot}7\times10^3$ dynes cm^{-3} for mid-continent basins we find that an average excess temperature of 240 °C would produce the observed subsidence upon cooling. This thermal anomaly would produce an uplift on about 800 m, well within the observed range of plateaux. Note that the coefficient of thermal expansion of continental lithosphere and the temperature anomaly cannot be determined independently if only the net amount of vertical displacement is known. A possible mechanism for producing this thermal anomaly is discussed below.

The sedimentary strata provide a record of the subsidence as well as episodic loads in the case of evaporities. Absolute ages were assigned to palaeontological ages using standard time scales (Harland, Smith & Wilcock 1965; Lambert 1971).

Atlantic margin of United States

It is useful to compare our results with the subsidence history of the Atlantic coast of the United States, as the geology there has been extensively studied. Although the general pattern of subsidence in Cretaceous and Tertiary times is compatible with the thermal contraction mechanism (Vogt & Ostenso 1967; Schneider 1969; Sleep 1971; Keen & Keen 1973), the sharp seaward termination of the marginal basin permits complications not relevant to closed basins, including seaward creep and thinning of the continental crust (Bott 1971) or the sedimentary prism (Cloos 1968) due to gravitational instability. Flexure due to sedimentary loads on the continental rise (Dietz 1963; Walcott 1972a) can be eliminated as the principal cause of subsidence on the east coast of the United States since the free air anomaly over the continental rise is too negative for an edge effect (Emery *et al* 1970). We note that along the Canadian Arctic coast (R. I. Walcott, 1975, private communication) and the Atlantic coast of Africa (Rabinowitz 1972), sedimentary loads on the continental rise cause a major part of the subsidence on the shelf and an excess positive anomaly near the edge of the shelf. Regional isostatic compensation for erosion in the nearby Appalachian mountains is an additional complication.

As progressively older beds crop-out landward on the Atlantic coast some information on the mechanical nature of the lithosphere can be made without knowing the detailed nature of the loading in the seaward parts of the basin. From Fig. 7 it is evident that the basin width a is about 150 km; thus α/a is about 2. Reasonable basin geometries result in Fig. 2 for values of α/a between 1·4 and 2·5. Flexural effects are not evident for smaller values of α/a. The gravity anomaly between the flanks and centre of the basin is excessive for large α/a, although quantification is difficult in this case. Using the measured α/a of 2 and rescaling t_e, we find that the

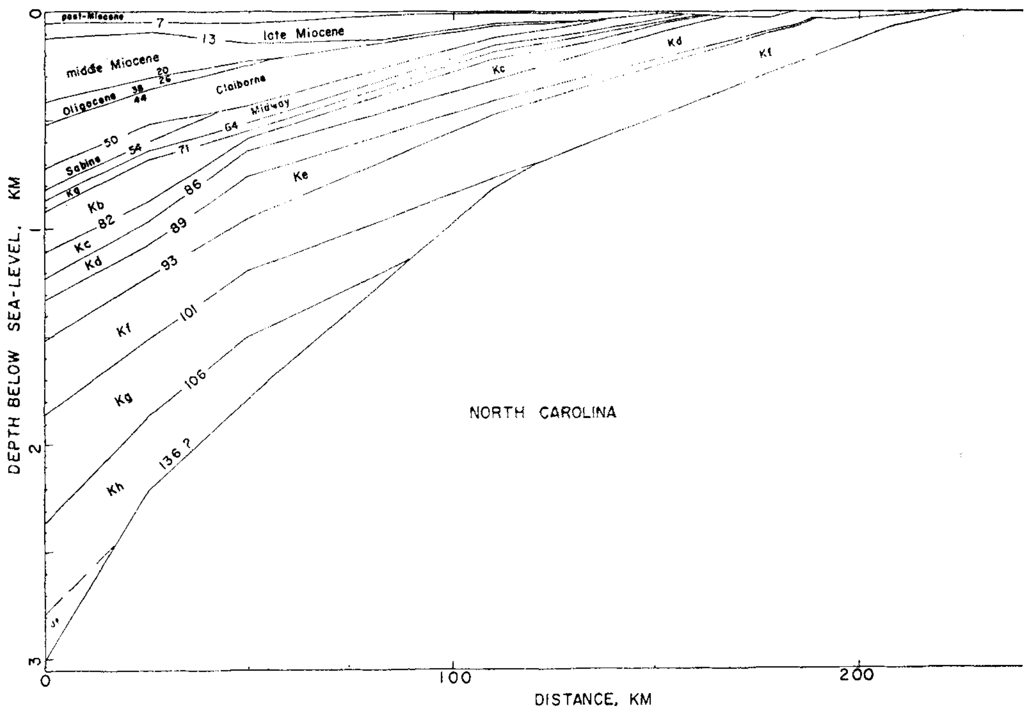

FIG. 7. Cross-section of the Atlantic continent shelf at Cape Hatteras (Data from Brown *et al.* 1972, cross-sections G–G′ and G′–G″, and fall line location from Maher 1965) is reversed for comparison with theoretical models. The shelf break is about 40 km from the left of the figure. The age of formation boundaries is given in millions of years. Where a significant unconformity is present the ages above and below are given.

acceptable results in Fig. 2 are equivalent to values of t_e between 0·25 My and 2·5 My, or viscosity between 2×10^{24} poise and 2×10^{25} poise, in general agreement with Walcott (1970a, b, c).

The exposure of older beds at local minima of subsidence as well as the flanks of the continental shelf basin provides further evidence that flexural effects, not merely random erosion due to eustatic lowerings of sea level, cause the landward outcrop of older beds. For example, the Florida arch of Ocala uplift has been a locally positive area since the Lower Cretaceous (Aplin & Aplin 1965, 1967). Neogene beds are absent over this arch and 0·4 km thick at well 3–2–25 at Cape May, New Jersey, although the base of the upper Cretaceous is about 1·2 km in both localities (Maher 1965). A different history of subsidence is thus necessary at these localities.

Michigan Basin

Rapid subsidence in the Michigan basin began after a major period of erosion in Middle Ordovician time. Older Cambrian and Lower Ordovician sediments underlie the basin, but the isopachs of these units do not have the characteristic bowl-shaped form of younger units but rather indicate a slightly-closed embayment of an area of general subsidence to the south (Catacosinos 1973). The Michigan basin is nearly free of later events, although some late Palaeozoic deformation is present (Ells 1969; Sloss 1963).

The outcrop pattern (Fig. 8) and cross-section (Fig. 9) of the Michigan basin indicate decreased basin width with younger age as predicted by our models. The

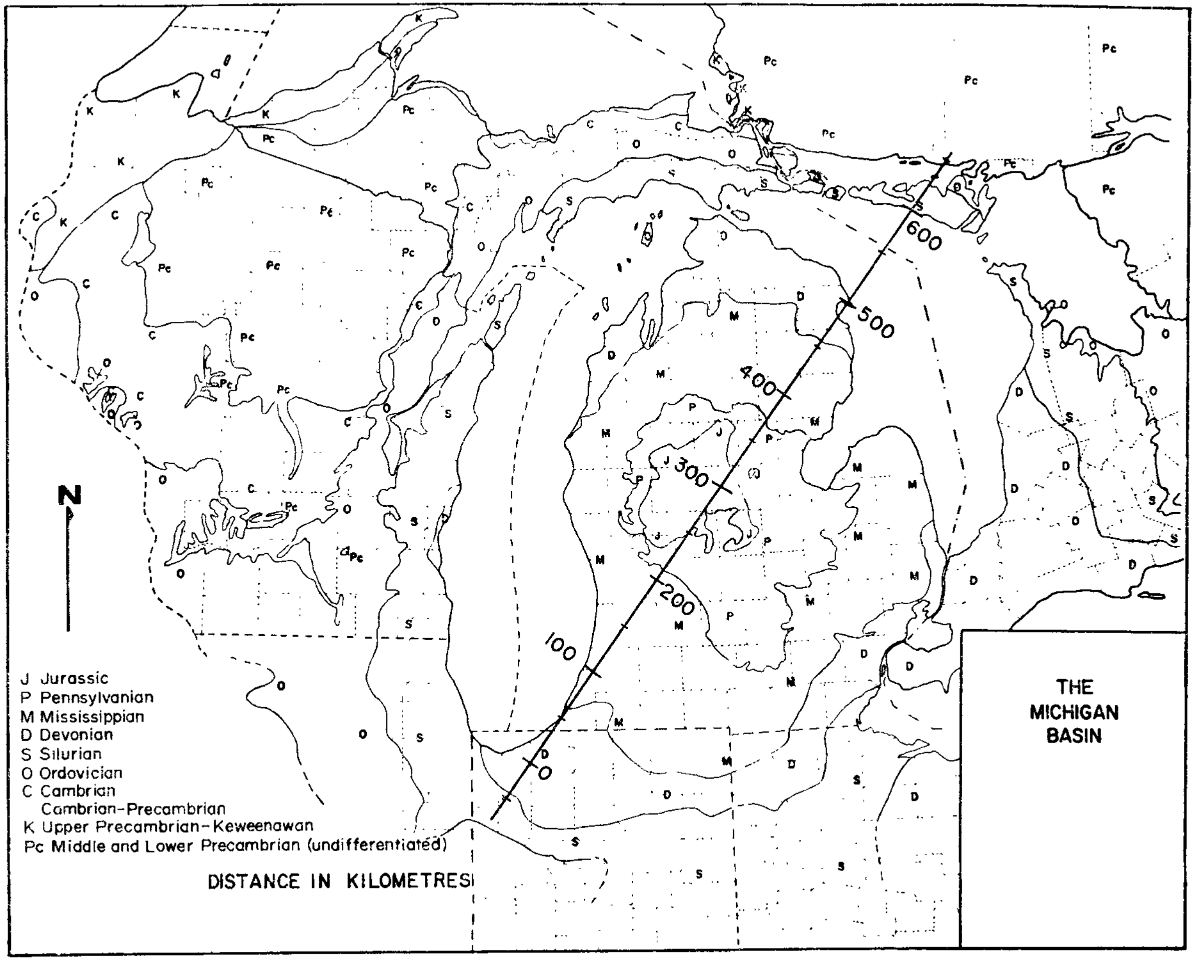

FIG. 8. Geologic map of the Michigan basin after Stonehouse (1969). Note the progressive outcrop of older beds away from the basin centre. The north-east striking path of the cross-section in Figs 9 and 10 was constructed to obtain maximum cover age by a path relatively perpendicular to the strike of the basin. (By permission of the Michigan Basin Geological Society.)

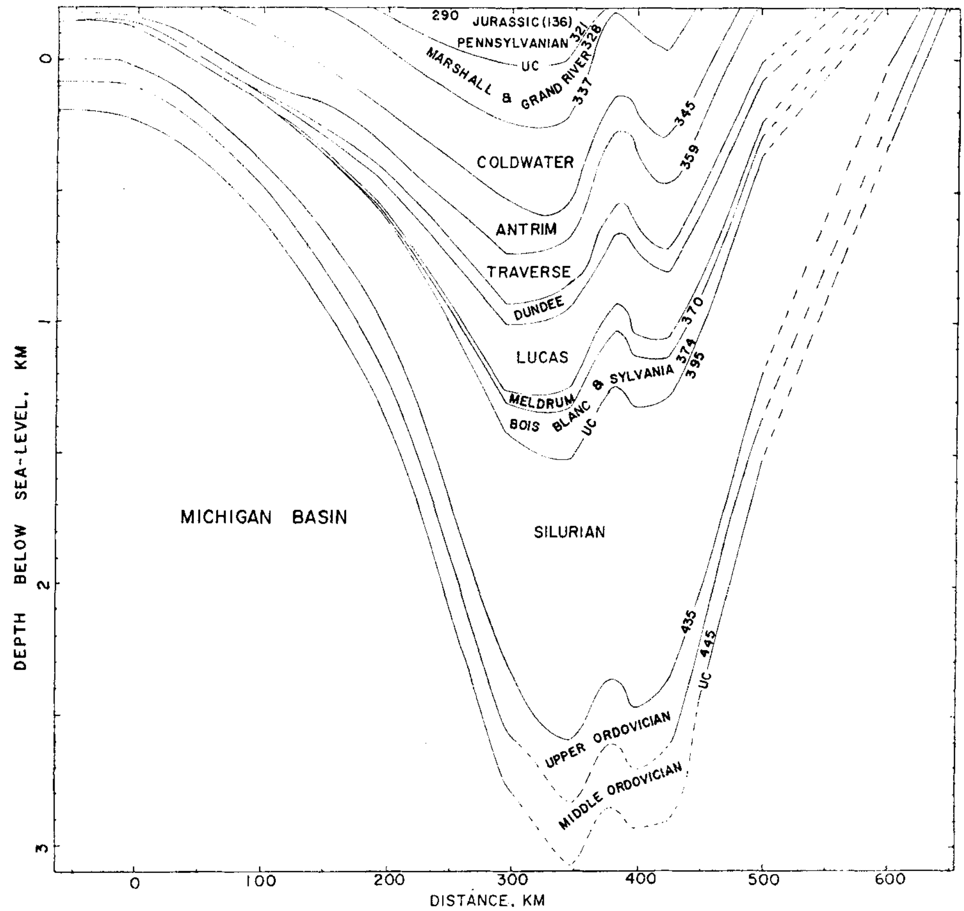

FIG. 9. Post-Sauk cross-section of the Michigan basin was constructed from published data (see Appendix). Formation boundaries are given absolute ages where possible. Major unconformities (Sloss 1963) are marked with 'UC' and the ages above and below given. Dashed lines indicate lack of data. No attempt was made to show the detail of minor structures, the distribution of Jurassic and later sediment, nor the position of the present surface. The anticline at 380 km is due to Late Palaeozoic tectonism. Devonian stratigraphic terminology after Gardner (1974).

models with α/a less than 1·4 do not show noticeable flexural effects (Fig. 3) and can thus be eliminated.

The gravity anomaly produced by the theoretical models is approximately equal to the attraction of the difference of the computed sediment thickness and the gaussian load causing subsidence. Models having a central thickness much less than unity can be eliminated if a corresponding gravity anomaly is not observed. The average free-air gravity anomaly in the central region of the Michigan basin is somewhere between +5 and +10 mgal (Walcott 1970c, 1972b). An additional subsidence of 0·2–0·4 km (6–12 per cent of the depth of the basin) for a density contrast of 0·6 g cm^{-3} would compensate the load indicated by the gravity anomaly. An α/a ratio in Fig. 3 of 2·5 would fit this observation. This value is an upper bound only however, since at least part of the anomaly is due to glacial erosion of the Great Lakes on the flanks of the basin and glacial deposition in the interior of the basin. A hundred or so metres of glacial topography could easily produce the positive anomaly.

In principle, the latest stages of basin development could be used to appraise flexural effects. It is not practical to do this because the latest stages of basin subsidence are prone to be obscured by eustatic changes and changes in local base level, and because small variations in the temperature at the base of the lithosphere may cause the late stages of subsidence to differ from the theoretical model. In the Michigan basin, the youngest Jurassic beds have been so extensively removed by glacial erosion that they nowhere outcrop. Some deformation of the basin by glacial loading is also conceivable. Another difficulty is that the youngest (Pennsylvanian and Jurassic) beds in the Michigan basin were deposited during major eustatic transgressions (Sloss 1963) and preserved by later subsidence of the centre of the basin. For example, the originally widespread occurrence of Pennsylvanian rocks is indicated by outliers in the Des Plaines, Illinois disturbance (Willman 1971). Evidence for Aptian (Middle Cretaceous) transgression has not been found in the Michigan basin, but if these rocks were similar to the Cretaceous sands at Quincy, Illinois (see Frye, Willman & Glass 1964), it is unlikely they would have been easily recognized beneath glacial or above Pennsylvanian deposits.

Late Palaeozoic* tectonism along north-west trending basement structure affected some parts of the present region of Michigan and Illinois (Ells 1969; Hinze & Merritt 1969; McGinnis 1966). Although the strike of the basin is parallel to the structures and some faulting with the downthrown side toward the interior of the basin may have occurred along these trends, the wavelength of a few kilometres and the discordant direction with respect to the basin of the structures precludes a simple explanation in terms of flexure and thermal contraction. It is likely that the state of stress beneath Michigan was extensively altered at that time. In any case, the locus of further subsidence shifted to the south-east after the Late Palaeozoic event by about 50 km (Fig. 9).

The progressive increase of bowl-shapedness of beds deposited during the times of more rapid subsidence provides a test of the flexural mechanism that is not sensitive to recent erosion within the basin or to Late Palaeozoic tectonism. In our models the ratio of amount of subsidence at $x = a$ over the amount of subsidence at $x = 0$ during the first 25 My of subsidence varied from about 0·37 at $\alpha =$ zero to 0·88 at $\alpha = 4{\cdot}5\,a$. The observed ratio for Middle and Upper Ordovician beds in the Michigan basin is about 0·62 corresponding to our model $\alpha/a = 1{\cdot}8$. This estimate, by using ratios of bed thicknesses, avoids uncertainties in the absolute time scale, but is sensitive to uniform additions to the thickness by eustatic effects and to variations in thickness due to depth of deposition. The thickness of Ordovician beds at the centre of the Michigan basin is also poorly constrained.

We can conclude from this discussion that the model with $t_e = 1$ My and $\alpha/a = 1{\cdot}4$ to 2·5 give acceptable results for the Michigan basin. As the observed α/a is about 1·7, viscosities for the lithosphere of 4×10^{24} poise to 4×10^{25} poise result (see equation (27)). This is in general agreement with our results for the Atlantic coast and Walcott (1970a, b, c).

Silurian salt in Michigan

In contrast with the normally bowl-shaped pattern of deposition in the Michigan basin, Silurian deposits below the lowest evaporite beds (Cataract and Casco carbonate, Fig. 10) are of relatively uniform thickness in the centre of the basin and somewhat thicker on the flanks. Although the beds on the flanks cannot be correlated

* In Michigan the time of formation of anticlinal structures is poorly constrained as Pennsylvanian strata are preserved only in the centre of the basin. These structures are not obvious in Mississippian isopachs and some truncation of the structures by Pennsylvannian strata is likely, but a good compilation of the data does not exist.

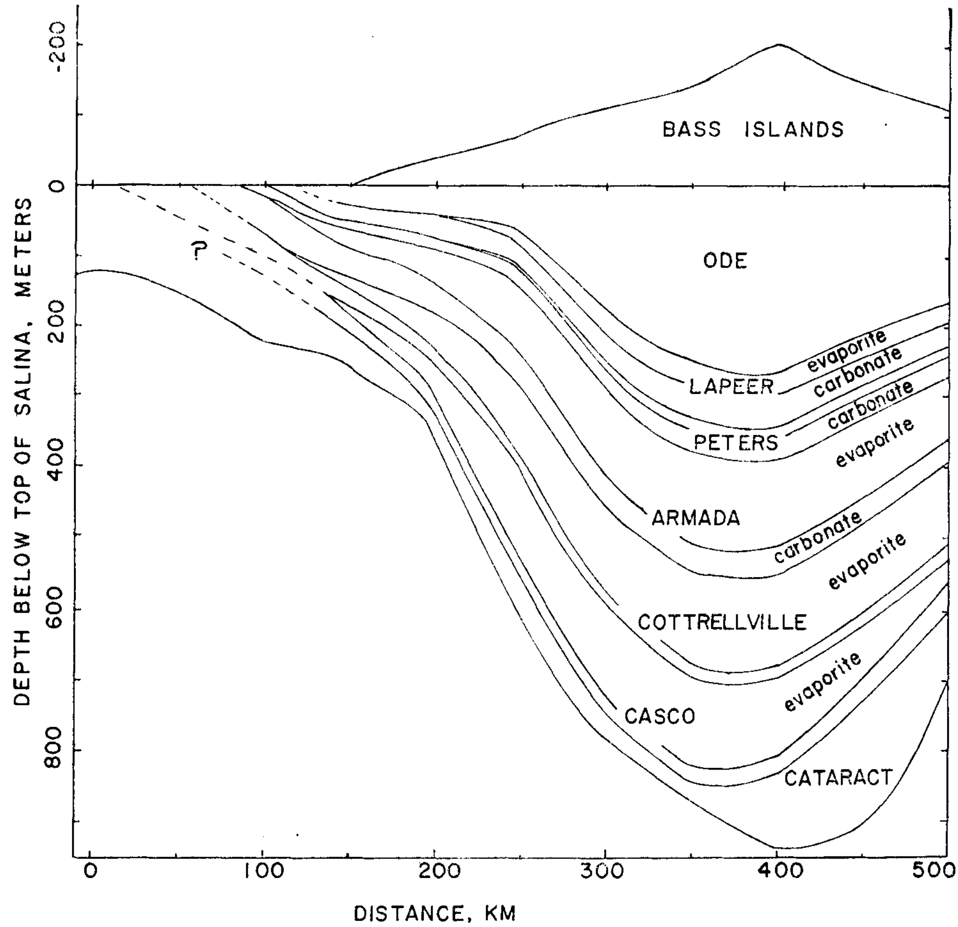

FIG. 10. Cross-section of the Michigan basin uses the top of the Silurian carbonate-evaporite sequence (Salina group) as a baseline for comparison with Figs 4, 5 and 6. The Ode formation consists of several carbonate-evaporite layers. Note the erosion on the flanks of the basin. See the Appendix for references. Terminology after Felber (1964).

with certainty with those in the interior (Briggs & Briggs 1974; Shaver 1974) and the absolute time scale for the Silurian period is poorly determined, this region is probably the most suitable for testing our mechanical model of an episodically filled basin. The evaporite beds within the basin form excellent time markers, although it is not certain that each evaporite filled the basin forming a palaeohorizontal (Elias 1964; compare McCrone 1964). The relevant observations to this problem can be summarized as follows (see Fig. 10).

During the early part of the Silurian land-derived clastic sediments ceased to be supplied to the basin. Subsequent carbonate deposition was thickest on the flanks of the basin where 200 m of reef deposits accumulated. About a third of these reefs are equivalent in time to the 30 m thick Casco carbonate in the basin interior (Mesolella *et al.* 1974). The nature of the Casco carbonate indicates that its reduced thickness is due to slow deposition in moderately deep water in the interior of the basin rather than local uplift (Mesolella *et al.* 1974; but compare Ehlers & Kesling 1962).

Thick evaporites with interbedded carbonates filled the basin. About ten evaporite carbonate cycles can be recognized (Ells 1962; Felber 1964; Allen 1974;

Mesolella *et al.* 1974). The individual layers of evaporites are up to 150 m thick near the centre of the basin and less than 30 m on the flanks (Allen 1974; Mesolella *et al.* 1974). Except in the far south-west corner of the basin where carbonate deposition continued, the reef complex on the flanks of the basin was covered by the Cottrellville and later evaporite layers (Briggs & Briggs 1974; Mesolella *et al.* 1974). Petrological and chemical evidence indicates that the evaporites at the centre of the basin precipitated in deep water, although the precise mechanism is uncertain (Sloss 1969; Schmalz 1969; Delwig & Evans 1969). Shallow water evaporites precipitated on the flanks of the basin (Kahle 1974). The entire carbonate-evaporite sequence may have accumulated in a short period of time at the end of the Silurian after the basin was starved. Alternatively the sequence may represent a nearly continuous filling interrupted by episodic sediment supply.

After the deposition of the evaporities, the Bass Islands carbonate was deposited in shallow water; extensive erosion caused by a major eustatic regression followed (Ells 1962; Sloss 1963, 1964, 1966; Sanford 1967; Gardner 1971). Part of the bowl-shapedness of the Silurian is due to preservation of the Bass Islands Formation only in the interior of the basin. Silurian evaporites and reefs were also exposed to this erosion. The evaporites were also subject to subsurface erosion by solution at this and later times. Middle Devonian and later sediments covered the post-Silurian surface, such that Silurian rocks are today exposed only on the flanks of the basin.

The geometry of the evaporites is not clearly compatible with large flexural effects from their load. There is a tendency for the evaporite to cover the reef on the flank but little indication that the initial deposits of evaporite extended greatly beyond their present extent. If the deposition was more or less continuous with equal intervals between evaporites, this would not be unexpected. Further subsidence does occur near the centre of the basin after evaporite deposition and erosion is more obvious on the flanks. This appears due mainly to a major eustatic transgression rather than flexural effects, which cannot uplift the entire basin at once. The models indicate that the observed later onlap over the flanks of the basin also cannot be due to flexural effects (Figs 4, 5 and 6). Any relaxation of flexural stress due to the evaporite load would enhance the erosion pattern in the observed direction. A useful estimate of relaxation time could not be obtained without additional knowledge of the amplitude of the eustatic regression.

It is also worth noting that the episodic manner in which individual salt beds were deposited could, in part, be explained as mechanical response to the load. A hundred metres of salt, a typical thickness of layer, would cause at least a few metres of downwarp of the flanks of the basin if the Earth had time to respond mechanically. The downwarp could re-open circulation into the basin, ending deposition. The response time of the asthenosphere to rapid loading (such as from glaciers) would control this downwarp. Renewed deposition of salt would begin when either eustatic variations, reef growth, or rebound of the basin flanks (in a time $\simeq t_e$) restricted circulation again. It is conceivable that this downwarp could be detected by careful study of the south-east flank of the basin where the Bass Islands Formation and underlying rocks are preserved.

Discussion and implications

As our models appear to be able to explain the uplift on the flanks and downwarp near the interior in the late stages of subsidence of mid-continent and Atlantic shelf basins, it is worth examining the implications of our results and the mechanisms of subsurface events which we have presumed to occur. The loading and heating necessary for later thermal contraction and subsidence and the rheological nature of the lithosphere are considered below.

Heat source and loading mechanism

For subsidence due to thermal contraction of the lithosphere to occur, it is necessary for the lithosphere to first become hot. Massive replacement of the cool uppermost mantle by hot asthenosphere is the most feasible mechanism, because conduction of heat from below would be as slow as the eventual cooling of the basin and require an extreme lateral temperature variation in the asthenosphere (Bott 1973; Sleep 1973). Unlike Atlantic continental margins, no obvious event related to the heating exists in the continental interior.

Sloss & Speed (1974) attribute the local subsidence of basins as well as eustatic changes to change and migration of the melt fraction of the asthenosphere. Before proposing an alternative mechanism based on intraplate stress, we note that:

(1) Intra-continental rifting and continental break-up occur episodically and contemporaneously throughout the world (Sloss 1972; Sloss & Speed 1974).

(2) These episodes coincide with eustatic regressions and oscillation and possibly with orogenies (Johnson 1971; Sloss & Speed 1974; see also Wise 1972).

(3) New basins begin subsiding after these episodes and transgressive periods are relatively free of intra-continental rifting (Sloss & Speed 1974).

(4) Eustatic transgressions are probably associated with rapid sea-floor spreading when much of the ocean floor is covered with young elevated crust. Regressions are probably associated with slow spreading (Valentine & Moores 1972; Flemming & Roberts 1973; Hays & Pitman 1973).

Inversions of observed plate velocities to determine the driving forces of plate tectonics have indicated that a critical velocity exists at which a downgoing slab would descend if it did not pull on the horizontal plates (Forsyth & Uyeda 1975; Tullis & Chapple 1973). More rapid descent would imply that the horizontal plates do work on the slab and that these plates are in horizontal deviatoric compression. If the descent rate were slower, both plates would be under horizontal tension. A relationship between slow sea-floor spreading and tensional rifting beneath continents could be produced by this mechanism. If minor rifting beneath continents can cause massive replacement of the uppermost mantle, we have a possible mechanism for heating the lithosphere beneath continents. If orogenies signal major changes in plate configuration, changes in spreading rate and interplate stress would also occur then.

The youngest continental margins and regions of rifting in continents are elevated as would be expected from thermal expansion. Some faulting and some uplift occurred before the Ordovician subsidence in the Michigan basin (McGinnis 1970). However, intra-plate stresses, the major element of our mechanism, have not been accurately determined for the present plates (Solomon, Sleep & Richardson 1975). It seems probable, though, that acceptable models will be found for the present and applied to past plate geometrics as they become known. It is not evident whether these results will be compatible with the hypothesis presented above.

It is necessary that the buoyancy of the continental crust be reduced in some way so that sediments can accumulate as thermal contraction occurs. This loading can result either from subareal erosion of material while the lithosphere beneath the basin is thermally expanded and elevated or by subcrustal processes (Sleep 1971, 1973; LePichon, Francheteau & Bonnin 1973, page 201). As in Kansas (Sleep 1971, 1973), presubsidence beds continue beneath the basin precluding erosion of as much material as later was deposited (Catacosinos 1973). The mechanism which loaded the crust in the Michigan basin during the Ordovician is not obvious. Subcrustal erosion of material off the base of the crust while the uppermost mantle was being replaced by hot material and intrusion of more dense material into the crust are possibilities (McGinnis 1970).

An elongate positive gravity anomaly, the mid-Michigan gravity high, is roughly centred and parallel with respect to the strike of the Michigan basin (Hinze 1963; Ells 1969; Hinze & Merritt 1969). As this anomaly probably is associated with Keeweenawan (1·2 By) igneous rocks (Muehelberger *et al.* 1967; Rudman, Summerson & Hinze 1965), it cannot be simply related to the load which caused Palaeozoic subsidence. It should also be noted that a basin does not overlie systematically this feature or the mid-continent gravity high.

Some relationship between the gravity high and the Michigan basin is likely, however, since faults associated with the mid-Michigan gravity high were active in the Late Palaeozoic (Ells 1969), and since Pennsylvanian and Jurassic subsidence is more centred on the gravity high than earlier subsidence. A more speculative possibility is that the production of the initial thermal anomaly may have been localized if the region of the mid-Michigan gravity high was thick, high density, isostatically compensated crust having greater lateral tension than surrounding regions during the Ordovician (see Hinze & Merritt 1969, for gravity interpretation; Artyushkov 1973, 1974). Also, thicker crust would kinematically be more likely to have its thickness reduced by subcrustal processes.

Rheology of lithosphere

It is of interest to compare the viscosity of the lithosphere of about 10^{25} poise which we obtained by considering cratonal basins with other possible estimates. As already mentioned, this value is in agreement with that obtained by Walcott (1970a, b, c) from the flexural response of the lithosphere to erosional, depositional, and volcanic loads of various ages. It is not surprising that we should find the processes inferred from Walcott's synoptic method to have been active through time. His estimate and ours are based on the same physical model of the lithosphere and the agreement between the methods should be taken for internal consistency rather than independent agreement.

The rheology of the lithosphere can also be estimated from the failure of continents to spread noticeably over the more dense mantle like oil over water (Artyushkov 1973, 1974). This tendency of regions of thick crust to be under tension can be observed in the patterns of Mesozoic continental break-up which follow Palaeozoic mountain belts (Hurley 1974). As horizontal stresses resisting this process are in the order of 100 bars, a one percent loss of continental freeboard would occur each 100 My if the relaxation time was 1 My ($\eta = 10^{25}$ poise). This much thinning has not taken place because Precambrian cratons exist at present sea level. Unless forces originating at plate boundaries maintain a compressive force in the lithosphere counter to this tensional force, we are faced with the problem of explaining why the lithosphere can deform under the few tens of bar stresses associated with basins and vertical loads, but not under horizontal 100 bar stresses. Artyushkov (1973, 1974) attributed this paradox to anisotropic behaviour of the lithosphere related to vertical faults.

Clearly the strains resulting from basin subsidence can be accommodated along high angle faults much more easily than uniform thinning and extension of the crust. The strains involved in basin subsidence, however, are much smaller than those needed to greatly thin the crust, and a ' transient ' or ' limited ' creep mechanism may produce creep in basins yet to be saturated with respect to horizontally directed stresses. Thus, before definitely attributing the relaxation of flexural stress to faults, it would be preferable to examine direct evidence for faulting.

A system of faults on the east coast of the United States has been hypothesized by Brown, Miller & Swain (1972). However, this arrangement is not evident in their original data (E. H. T. Whitten 1975, private communication), and seems to have been at least partially preordained, as their tectonic theory presupposes a conjugate system of faults. The extent to which faulting occurred during subsidence on the Atlantic continental shelf cannot thus be inferred at present.

For the Michigan basin, it can be definitely concluded that down-to-basin faults accommodated at least some of the deformation within the lithosphere. These basement faults, which were probably originally active during Keeweenawan times, are associated with north-west striking gravity and magnetic anomalies and petroleum producing trends parallel to the axis of the basin (Ells 1969; Hinze & Merritt 1969). The history of faulting is difficult to work out, however, since basement faulting may produce monoclinal flexure in the overlying sediments and since movements in ductile salt beds may amplify the deformation. More detailed studies are needed to quantify this point.

Another estimate of the rheology of the lithosphere can be obtained from the observation that thermoelastic stresses and stress due to movement of plates over an elliptical earth probably relax more or less isotropically without creating large stress. These mechanisms have been proposed as the cause of seamount chains and African rift valleys (Oxburgh & Turcotte 1974; Turcotte & Oxburgh 1973). Although a convincing case might be argued if only one feature is considered at a time, the overall distribution of seamounts is not consistent with a thermoelastic or elliptical stress mechanism. The thermoelastic mechanism would imply a consistent alignment of seamount chains with local magnetic anomaly rather than the observed consistent alignment within a plate. Stress due to the non-spherical shape of the Earth would be greatest on moving plates at mid-latitudes contrary to the observed location of seamounts both north and south of the equator on the Pacific plate (Solomon & Sleep 1974; Solomon *et al.* 1975).

If we suppose that thermoelastic stresses must remain below the level of intraplate stress, that is, less than about 100 bars, we can obtain a maximum estimate of the viscosity with respect to these stresses. Thermoelastic strains should be in the order of the linear thermal expansion of the lithosphere and mid-ocean ridges, about one per cent. Remembering that the lithosphere at ridges cools on about a 50 My time scale and that 100 bars is equivalent to about 0·01 per cent strains, we find that the thermoelastic stress must relax on a 0·5 My time scale ($\eta = 5 \times 10^{24}$ poise), in line with our estimate from the deformation of cratonal basins.

The observation answers in part the question as to whether anisotropy due to vertical faults or a limit to the amount of creep which can occur in the lithosphere is the explanation for the different behaviour under flexural and horizontal stresses. Both flexural stress in mid-continent basins and thermoelastic stress relaxations require only small strains while large strains are required to thin the crust by spreading. It should also be noted that thermoelastic stress relaxation must be distributed even on the level of grain size, since differences in thermal expansion coefficients between likely minerals and the level of anistropy are subequal though smaller than the average expansion (Skinner 1966). A saturatable creep mechanism for the lithosphere would also explain why deformation due to large loads such as seamount chains apparently ceases after 10 My. (Watts & Cochran 1974; Watts & Talwani 1974, 1975), and why mountain belts and other local heterogeneities in crustal thickness can persist through geological time.

An alternative model, which would produce creep saturation, consists of a thin, highly viscous surface layer existing in the uppermost part of the lithosphere. The layer would not greatly affect the response of the lithosphere to distributed loads, but would preclude large strains from either concentrated surface loads or horizontal stress (R. I. Walcott & J. P. Foucher 1975, private communication).

The lithosphere must be able to undergo large irreversible strains where slabs plunge into the mantle during continental breakup. High viscosities such as we have discussed would clearly preclude these processes, unless another creep mechanism were available. At trenches we must appeal to non-linear weakening of the material at stresses higher than the mid-plate stresses (Sleep 1975). Intrusions of hot low-viscosity material along brittle cracks clearly play a role at ridges. More work is needed to clearly define the rheology of the lithosphere including direct determination of when strain concentrates on faults and when it is distributed throughout the lithosphere.

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Appendix

Compilation of Michigan basin cross-sections

In order to compare our theoretical model with observations it was necessary to construct a cross section of the Michigan basin. The stratigraphy developed for the outcrop areas on the flanks of the basin is for the most part difficult to correlate with the sub-surface, as facies changes occur and other non-outcropping units appear in the region of maximum subsidence. Some of the published maps ended at the state line necessitating correlation with Indiana maps drawn by other investigators. The isopachs of individual units, however, are summed to the proper interval between available structural contours.

For purposes of reference we present a listing of the sources used and a correlation of the units to the European or North American time scales. Layers identified as time units rather than time transgressive rock units were used as much as possible.

1. Jurassic—not shown on section
   age: Kimmeridgian (Ells 1969, p. 62)

2. Pennsylvanian (Cohee 1951)
   age: Pottsville to Conemaugh (Ells 1969, p. 62)

3. Marshall & Grand River (Cohee 1951)
   age: Osagian to Meramecian, Mississippian (Ells 1969, p. 62)

4. Coldwater (Cohee 1951)
   age: Kinderhookian, Mississippian (Ells 1969, p. 62)

5. Antrim (Cohee 1947, 1951; Lineback 1968; Shaver *et al.* 1971) includes Ellworth shale, Sunbury shale, Berea sandstone Bedford shale and Antrim shale (Ells 1969, p. 62). The contacts within this interval are time transgressive (Sanford 1967; Asseez 1969; Lineback 1968; Dewitt 1970)
   age: Frasnian, Devonian to lowest Mississippian (Sanford 1967)

6. Traverse (Gardner 1974; Shaver *et al.* 1971)
   age: Givetian, Devonian (Sanford 1967)

7. Dundee (Gardner 1974) includes Rogers City Limestone (Ells 1969, p. 62)
   age: upper Eifelian, Devonian (Sanford 1967)

8. Lucas (Gardner 1974; Shaver *et al.* 1971) includes Anderdon formation and Richfield zone (Ells 1969, p. 62)
age: middle Eifelian, Devonian (Sanford 1967)

9. Meldrum (Gardner 1974) equals Black lime zone of Amherstburg formation (Ells 1969, p. 62) and ' Filer sandstone '
age: lower Eifelian, Devonian (Sanford, 1967)

10. Bois Blanc and Sylvania (Gardner 1974)
age: Emsian and lowest Eifelian (Sanford 1967)
includes Garden Island formation (Ells 1969, p. 62) of Siegenian, Devonian age (Sanford 1967)

11. Silurian (Fisher 1969; Shaver *et al.* 1971) for Fig. 9

(a) Bass Islands (unpublished well logs, David Larue, analyst)

(b) Ode (Felber 1964; Allen 1974) equals G unit and F unit except for lowest F salt of Ells (1962)

(c) Lapeer (Felber 1964; Allen 1974; Ells 1962) equals lowest F salt and E unit of Ells (1962)

(d) Peters (Felber 1964; Allen 1974; Ells 1962) equals C and D units of Ells (1962)

(e) Armada (Felber 1964; Allen 1974; Ells 1962; Mesolella *et al.* 1974; Shaver et al. 1971) equals A-2 carbonate and B units of Ells (1962)

(f) Cottrellville (Felber 1964; Ells 1962; Mesolella *et al.* 1974; Shaver *et al.* 1971) equals A-2 evaporite and A-1 carbonate of Ells (1962)

(g) Casco (Felber 1964; Ells 1962; Mesolella *et al.* 1974; Shaver *et al.* 1971) Shaver *et al.* 1971) equals A-1 evaporite and in sub-surface Niagaran of Ells (1962)

(h) Cataract (Fisher 1969) obtained from Silurian isopach minus above units

12. Upper Ordovician (Cohee 1948; Hinze & Merritt 1969) equals Richmond age

13. Middle Ordovician (Cohee 1948) equals Trenton and Black River ages