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SOME REMARKS ON THE DEVELOPMENT OF SEDIMENTARY BASINS

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A simple model for the development and evolution of sedimentary basins is proposed. The first event consists of a rapid stretching of continental lithosphere, which produces thinning and passive upwelling of hot asthenosphere. This stage is associated with block faulting and subsidence. The lithosphere then thickens by heat conduction to the surface, and further slow subsidence occurs which is not associated with faulting. The slow subsidence and the heat flow depend only on the amount of stretching, which can be estimated from these quantities and from the change in thickness of the continental crust caused by the extension. The model is therefore easily tested. Preliminary investigations of the Great Basin, the Aegean, the North Sea and the Michigan Basin suggest that the model can account for the major events in their evolution.

1. Introduction

The general acceptance of plate tectonics and of the associated thermal models of plate creation has successfully accounted for the major horizontal and vertical motions of the ocean floor. No comparable progress has occurred in our understanding vertical movements in continental regions, even though the development of several major basins is now known in some detail. The North Sea [1,2] illustrates some of the problems involved. It contains more than 4 km of sediments which have accumulated in a subsiding basin during the Cretaceous and Tertiary. The Tertiary sediments along are more than 2 km thick over a large part of the sea and those younger than Paleocene have suffered almost no deformation. There is general agreement that the whole region is underlain by continental crust, and that the only deep-water sediments are those deposited in the grabens. Since the gravity anomalies are small the subsidence must be compensated. Two explanations of these and similar observations in other basins have been widely discussed for some time. The older depends on the Moho being a phase change. The sedimentary loading increases the pressure at depth and hence the Moho migrates upward. Collette [3] has proposed such a model for the North Sea and O'Connell and Wasserburg [4] have worked out the general behaviour of such a model. These ideas have not been generally accepted, principally because few geophysicists now believe that the Moho is a phase change. The other class of models [5–8] depend on thermal contraction to produce the subsidence and closely resemble the thermal models which have been so successful in the oceans [9]. Though these models account rather well for the time history of the subsidence, there is a major space problem which Sleep [5] emphasized. If the temperature of the continental lithosphere is raised by some means, the surface is elevated, gradually subsiding to its original position as the lithosphere cools. Hence a surface originally at sea level finally returns to sea level. To produce a basin without crustal stretching material must be removed on a large scale by erosion when uplift occurs. There is no evidence that 2 km or more of rock was removed from the North Sea after the block faulting during the Jurassic and Cretaceous. The little erosion that did occur is local and associated with the horsts produced by normal faulting [10, p. 190]. The stratigraphy of this and other sedimentary basins show that their history starts with extensive normal faulting and subsidence. Only local horsts are elevated and eroded. Haxby et al. [8] avoided this difficulty by appealing to phase changes, but added little to the previous discussion of this question.

The other difficulty is the mechanism which heats the continental lithosphere. Haxby et al. [8] offered an imaginative model based on mantle diapirs, which are supposed to be able to intrude and replace the lower part of the lithosphere with hot rock without producing major deformation at the surface. In oceanic regions no heat flow anomalies have yet been discovered which require variations in temperature in the mantle beneath the lithosphere. Both the heat flow anomalies associated with ridges [9] and with back arc spreading [11] can be produced by a simple plate model underlain by an isothermal mantle. Furthermore the oceanic heat flow anomalies are larger and more extensive than those found in continents. It would therefore be surprising if continental, rather than oceanic, heat flow anomalies required the existence of local heat sources beneath the lithosphere.

Both the space and the heating problem are avoided if the basin is produced by stretching continental crust over a large region. Such a model has often been used to account for the normal faulting and crustal thinning observed in rifted regions [12,13], but the thermal consequences of the stretching have received little attention. An extensional model of this type can account for the present deformation and heat flow in the Aegean Sea region [14], though, since the deformation is still occurring, no undeformed sedimentary basin has yet developed. Unlike the models of Voight [12] and Makris [15], no thermal anomalies associated with basin formation are required in the asthenosphere. The prupose of the development below is to examine the surface flow and subsidence produced by arbitrary amounts of stretching. The results are then compared with the evolution of several basins and used to suggest how the stretching model can be tested.

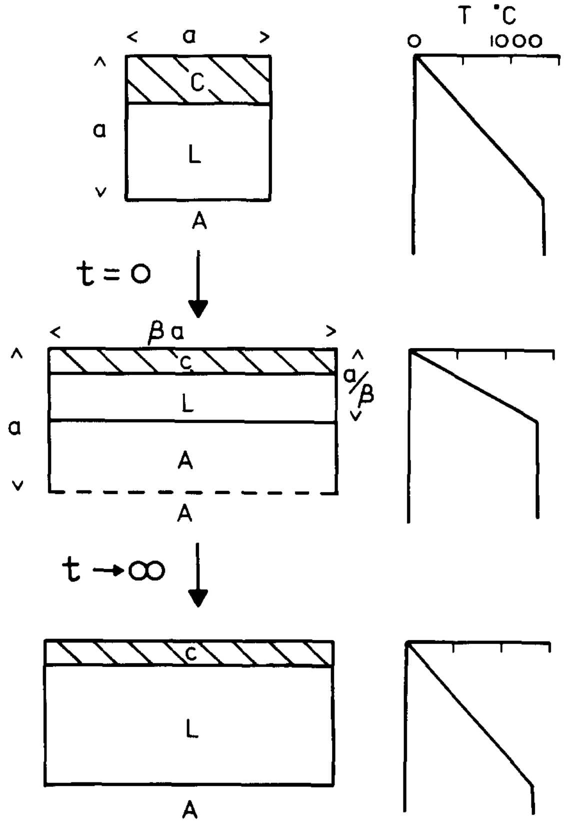

Fig. 1. Sketch to show the principal features of the subsidence model. At time $t = 0$ a piece of thermally equilibrated continental lithosphere is extended by β. Since the temperature of the material remains unchanged during the extension, isostatic compensation causes upwelling of hot asthenosphere. Cooling of this hot material produces subsidence as the temperature perturbation decays. Continental crust is assumed to be conserved during extension and its radioactivity neglected. The discontinuity in the temperature gradient between the lithosphere and the asthenosphere is an artifact of the model which could be removed by considering the details of the convective heat transport in this region [27]. However, the heat flux and subsidence would be little affected.

2. Model calculations

The simplest stretching model is illustrated in Fig. 1. At time $t = 0$ a unit length of continental lithosphere is suddenly extended to a length β, causing upwelling of hot asthenosphere. The resultant thermal perturbation gradually decays, producing subsidence. Isostatic compensation is preserved throughout. The simplest model ignores the radioactivity of continental rocks and assumes that the temperature at a depth corresponding to the initial thickness of the lithosphere is fixed. More realistic models can easily be examined by the same techniques, but the algebra obscures the simple physical ideas involved.

Since the lithosphere is isostatically compensated

TABLE 1

Values of parameters used (mostly taken from Parsons and Sclater [9])

a	=	125 km
ρ_0	=	3.33 g cm^{-3}
ρ_c	=	2.8 g cm^{-3}
ρ_w	=	1.0 g cm^{-3}
α	=	3.28×10^{-5} °C^{-1}
T_1	=	1333°C
τ	=	62.8 My
kT_1/a	=	0.8 μcal cm^{-2} s^{-1}
E_0	=	3.2 km

both before and after extension, there is an initial subsidence S_i given by:

$$S_i = \frac{a\left[(\rho_0 - \rho_c)\frac{t_c}{a}\left(1 - \alpha T_1 \frac{t_c}{a}\right) - \frac{\alpha T_1 \rho_0}{2}\right]\left(1 - \frac{1}{\beta}\right)}{\rho_0(1 - \alpha T_1) - \rho_w} \tag{1}$$

where a is the thickness of the lithosphere and t_c the initial thickness of the continental crust, ρ_0 the density of the mantle, ρ_c that of the continent both at 0°C. ρ_w is the density of seawater, α the thermal expansion coefficient of both the mantle and the crust and T_1, the temperature of the astenosphere. The surface of the continent is taken to be at or below sea level, and continental crust is assumed to be conserved. The sign of S_i depends on t_c and is independent of β. Using values for the quantities in (1) in Table 1, taken from Parsons and Sclater [9], S_i is positive if $t_c \gtrsim 18$ km. Hence land areas will subside but regions with thin crust can be elevated by the stretching, though not sufficiently to emerge above sea level. It is of course possible that uncompensated islands may be produced during the extension by block faulting as has happened in the Aegean, or that vulcanism may cause the volume of the continental crust to increase. These processes can elevate part or all of a stretched basin above sea level during extension which would otherwise sink. The stretching increases the heat flow by β at $t = 0$ if it occurs instantaneously. After the extension the temperature variation is:

$$T = T_1\,, \qquad 0 < \frac{z}{a} < \left(1 - \frac{1}{\beta}\right)$$

$$= T_1\beta\left(1 - \frac{z}{a}\right), \qquad \left(1 - \frac{1}{\beta}\right) < \frac{z}{a} < 1 \tag{2}$$

where z is measured upwards from the base of the lithosphere before extension. To determine the subsidence and heat flow as functions of time we must solve the one-dimensional heat flow equation:

$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2} \tag{3}$$

where κ is the thermal diffusivity, with boundary conditions:

$$T = 0\,, \qquad z = a$$

$$= T_1\,, \qquad z = 0 \tag{4}$$

and initial conditions (2). This is most easily done by Fourier expansion, and is very similar to a solution given by Lubimova and Nikitina [16] for the temperature structure in an oceanic plate. The solution for T is:

$$\frac{T}{T_1} = 1 - \frac{z}{a} + \frac{2}{\pi}\sum_{n=1}^{\infty} \frac{(-1)^{n+1}}{n}\left[\frac{\beta}{n\pi}\sin\frac{n\pi}{\beta}\right] \times \exp\left(\frac{-n^2 t}{\tau}\right)\sin\frac{n\pi z}{a} \tag{5}$$

where:

$$\tau = \frac{a^2}{\pi^2 \kappa} \tag{6}$$

As $\beta \to \infty$ the term in square brackets $\to 1$ for all n and the solution is that for an oceanic ridge model when all the heat is conducted vertically [9]. The surface heat flux, F, is easily obtained from (5):

$$F(t) = \frac{kT_1}{a}\left\{1 + 2\sum_{n=1}^{\infty}\left[\frac{\beta}{n\pi}\sin\frac{n\pi}{\beta}\right]\exp\left(-\frac{n^2 t}{\tau}\right)\right\} \tag{7}$$

The elevation, $e(t)$, above the final depth to which the upper surface of the lithosphere sinks is:

$$e(t) = \frac{a\rho_0\alpha T_1}{\rho_0 - \rho_w}\left\{\frac{4}{\pi^2}\sum_{m=0}^{\infty}\frac{1}{(2m+1)^2} \times \left[\frac{\beta}{(2m+1)\pi}\sin\frac{(2m+1)\pi}{\beta}\right]\exp\left(-(2m+1)^2\frac{t}{\tau}\right)\right\} \tag{8}$$

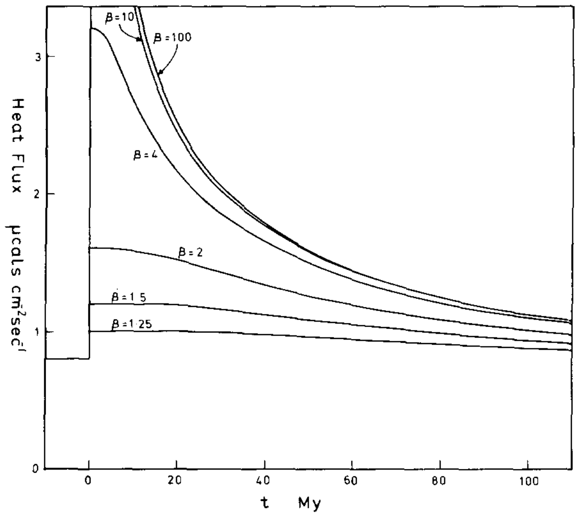

Fig. 2. Heat flux as a function of time for various values of β, obtained from equation (7).

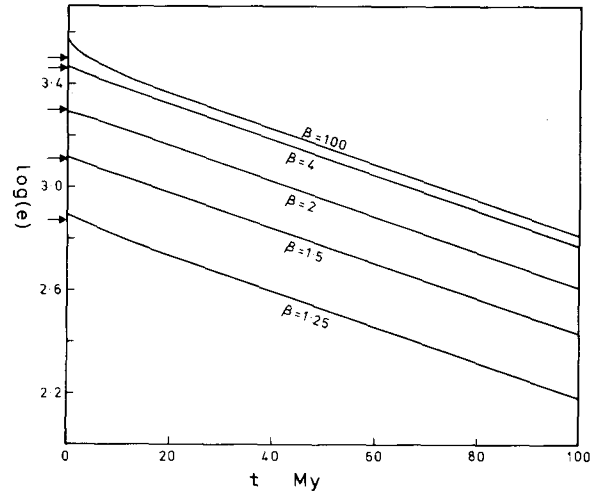

Fig. 3. $\mathrm{Log}_{10}[e(t)]$, where $e(t)$ is in metres, as a function of time for various values of β obtained from equation (8). The arrows mark the positions where straight lines fitted to the curves for values of $t \geqslant 20$ My intersect the $t = 0$ axis.

The heat flux is a function of time for various values of β in Fig. 2 shows a strong dependence on β for time less than about 50 My if β is between 1 and 4. However the heat flux is insensitive to β when β is large because almost all the region between $z = 0$ and $z = a$ is replaced with asthenosphere during extension, and the thin remnant of the original lithosphere has little influence. Extension increases the heat flux by a factor β. After a time which depends on the thermal time constant of the stretched lithosphere, τ/β^2, the heat flux starts to decrease. The behaviour at large times, $t \gtrsim 30$ My, can be described by the first term of the summation in (7):

$$F(t) = \frac{kT_1}{a}\left[1 + 2r \exp\left(-\frac{t}{\tau}\right)\right] \tag{9}$$

where $r = (\beta/\pi) \sin(\pi/\beta)$ (10)

is the fraction by which the time-dependent part of the heat flux is reduced below the ridge model. Since $\beta > 1, 0 < r \leqslant 1$. Hence when $t << \tau/\beta^2$ the ratio of the heat flux after stretching to that before gives β directly.

At times between 0 and 30 My β can only be obtained from F by using (7), but for later times (9) is sufficient. When $t \gtrsim \tau$ the heat flow anomaly is small and will not be easy to observe, and for values of $\beta \lesssim 1.5$ the anomaly will be difficult to observe at all times.

The behaviour of the elevation anomaly is different. Fig. 3 shows $\log_{10}(e)$ as a function of t for various values of β. The curves are almost straight lines when $\beta < 4$ for all values of t. The reason for this is clear from (8): for such values of β the second term in the summation is very small and:

$$e(t) \simeq E_0 r \exp(-t/\tau) \tag{11}$$

where:

$$E_0 = \frac{4a\rho_0\alpha T_1}{\pi^2(\rho_0 - \rho_w)} \tag{12}$$

Parsons and Sclater [9] give a value of 3.2 km for E_0. When β is large and $r \to 1$ the corresponding expression to (11) is valid only for $t \gtrsim 20$ My. The subsidence since extension, S_t, is sometimes more easily measured than e:

$$S_t = e(0) - e(t) \tag{13}$$

The total subsidence S is the sum of S_t and S_i.

Following Parsons and Sclater [9], S_t is shown as a function of $\sqrt{t}$ in Fig. 4. An approximate expression for S_t, σ_t, may be obtained from (11):

$$\sigma_t = E_0 r[1 - \exp(-x^2/\tau)] \tag{14}$$

where:

$$x^2 = t \tag{15}$$

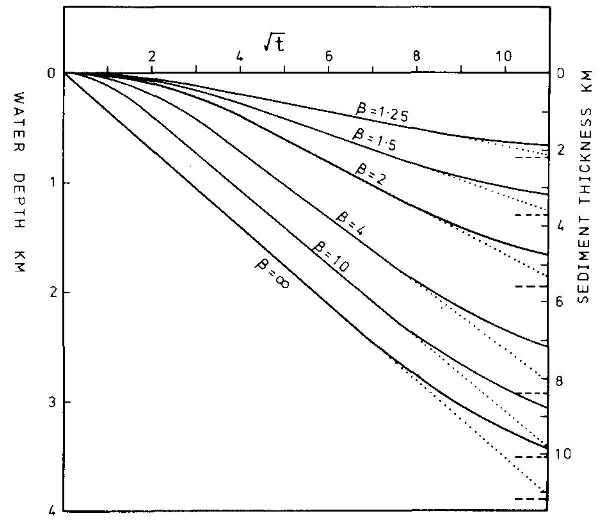

Fig. 4. Subsidence S_t as a function of $\sqrt{t}$, where t is in My, on the left, sediment thickness on the right (see text). The continuous curves are obtained from equation (13), the straight dotted lines were obtained by least squares fitting between $\sqrt{t}$ = 4 and $\sqrt{t}$ = 7. The horizontal dashed lines show the asymptotic subsidence for each curve.

TABLE 2

Comparison between full and approximate expressions

	β	$e(0)$	m	C	t for 5% deviation in My
F	1.25	781	80.1	−118	68.04
A	1.25	740	80.1	−158	
F	1.50	1300	140	−273	70.12
A	1.50	1309	142	−279	
F	2.0	1950	213	−460	71.49
A	2.0	2015	218	−429	
F	4.0	2930	307	−520	68.82
A	4.0	2850	308	−607	
F	10.0	3520	340	−285	66.84
A	10.0	3114	337	−663	
F	100.0	3860	347	−6	66.40
A	100.0	3165	343	−674	
F	00	3890	347	19	66.40

The subsidence is described by $S_t = m\sqrt{t} + C$. m and C for the full theory, marked F, were obtained by the method described in the text, those for the approximate theory, marked A, from equation (16). All parameters describe a basin containing no sediment.

All curves for finite β are tangential to the $\sqrt{t}$ axis at $t = 0$, are approximately straight lines between 16 and 60 My, then trend to an asymptotic value $e(0)$. The behaviour near the origin is unlikely to be observable, but both the slope of the straight-line portion and the asymptotic value can be directly measured, and it is useful to obtain simple expressions for both. The straight line which best fits (14) is the tangent where $d^2\sigma_t/dx^2 = 0$, or where $x^2 = \tau/2$:

$$\sigma_t = E_0 r \sqrt{\frac{2}{e\tau}}\sqrt{t} - E_0 r\left(\frac{2}{\sqrt{e}} - 1\right) \tag{16}$$

Table 2 shows the gradient and intercept obtained by fitting a straight line to (13) between values of $\sqrt{t}$ of 4 and 7, marked F, and those obtained from (16), marked A. The gradient agrees excellently for all values of β; the agreement for intercept values is good only when $\beta \lesssim 4$. The asymptotic value of the subsubsidence is $e(0)$, or approximately $E_0 r$, shown in Fig. 3.

When $t \gtrsim \tau$ the subsidence curves in Fig. 4 deviate from straight lines. The time at which this deviation exceeds 5% of the remaining subsidence is given in Table 2, and shows a weak dependence on β. An approximate expression obtained from (14) and (16) is independent of both E_0 and r, and hence β, depending only on τ. Clearly β should not be estimated from this part of the curves.

Most basins whose basement is continental crust contain large thicknesses of sediment, and hence it is the sediment thickness rather than the subsidence which is directly observed. To obtain the subsidence the sediment load must be removed using either an Airy model or a flexure model [7] for the isostatic compensation. Since the sediment density is variable and compaction also must be taken into account, such compensation must be carried out separately for each basin. For this reason the discussion above has been concerned with the subsidence of an empty basin. For reference, however, the corresponding values for the thermal subsidence of a basin initially at sea level completely filled with sediment of density 2.5 g cm^{-3} have been added to the right of Fig. 4. This subsidence will be increased if S_i is not zero and the basin is filled to sea level.

The other difficulty is the influence of continental radioactivity, which affects the heat flux more than the temperature distribution. Observations from a number of continental wells indicate that the radioactive elements involved are strongly concentrated towards the surface, and therefore that the influence on the temperature distribution throughout most of the continental lithosphere is small. If the thickness of the layer containing most of the heat-generating elements is ignored, the contribution to the heat flux is easily determined. If the heat flux from this source is F_c before stretching it is reduced to F_c/β after extension, and the steady-state heat flux is reduced. This decrease may be partly concealed by heat generation in the sediments deposited in the basin.

This discussion of a simple thermal model for extension shows that both the heat flux and the subsidence depend on the amount of extension through a quantity r given by (10). In the first 30 My after extension the heat flux is strongly affected by the amount of extension, whereas at later times the subsidence provides a better estimate. If the subsidence is plotted against $\sqrt{t}$ simple expressions describe both the gradient of the straight-line portion (equation (16)) and the asymptotic value (equation (14)) of the curves. The only variable is β, the amount of extension.

The discussion below is concerned with the development of sedimentary basins. The model is, however, also relevant to the temperature structure and shape of ridge axes, where rapid stretching continuously thins the plate. The expressions above should allow the thermal structure of the plate at ridge axes to be estimated directly from the bathymetry.

3. Geological and geophysical observations

The most obvious objection to the model discussed in the last section is that the large amounts of extension required to produce the observed subsidence have not been described. This objection is probably the reason why the model discussed above had not previously been proposed. Two regions, the Great Basin and the Aegean Sea, presently undergoing extension, have been studied in some detail using a variety of geological and geophysical techniques. In the Aegean region there are a number of normal faults which have steep dips at the surface, but earthquake fault plane solutions [14] show that the dips decrease with depth. Similar listric faults have been proposed by Murawski [17] to account for focal mechanisms in the Rhine Graben. In the Canadian Rocky Mountains Bally et al. [18] have argued that the surface expression of normal faults is only compatible with the seismic reflection records if their dip decreases with depth. Movement on such curved faults produces rotation of the sediment horizons, which is often easily visible on reflection records, especially after these have been converted to depth sections. The dip of faults themselves is often hard to determine partly because of the large vertical exaggeration of conventional displays, but the characteristic rotation is commonly observed on reflection profiles in regions where extension has occurrred. The curvature of listric faults is important because it prevents any accurate estimate of the extension from surface observation of dip and throw of the faults. Hence major extension cannot easily be estimated by surface mapping. Probably the best method of determining the extension is from the crustal thickness obtained from seismic refraction, especially if adjacent unextended crust of similar basement geology is available for comparison. In the Aegean region seismic refraction has been carried out by Makris [19] and Makris and Vees [20] in both the stretched and unstretched areas, and suggests extension by a factor of about two ($\beta \simeq 2$). This value is compatible with the heat flow observations of Jongsma [21]. Any volcanic additions to the continental crust during extension will cause a decrease in this estimate of β, but since the added thickness is unlikely to exceed the thickness of the oceanic crust, or about 5 km, the resulting error is unlikely to be important unless $\beta \gtrsim 4$.

In the Great Basin of the western U.S.A. recent mapping has demonstrated the existence of a number of large shallow-angle normal faults [22]. Seismic reflection observations (Snelson, person communication) show that similar faults exist at depth in other parts of the basin. It therefore appears probable that the extension in the Great Basin is considerably greater than commonly estimated from surface mapping.

Because the extension is very recent and still continues, the thermal contribution to the subsidence of both the Aegean and the Great Basin is small. In older baisns, such as the North Sea, great thicknesses of sediments have accumulated. The center of the North

Sea is underlain by a thickness of at least 3 km of sediment, which has accumulated since the Upper Cretaceous [1]. Some of these Tertiary sediments accumulated in the deep water of the median graben [2,23] and therefore cannot be used to estimate the extension directly. However, at least 2 km of shallow-water clastic deposits have accumulated, giving an approximate value of β estimated from Fig. 4 of 1.5. This is somewhat larger than Collette's [3] estimate from gravity observations. The seismic refraction results obtained by Sornes (see Wilmore [24]) do not shown any crustal thinning, but this may be a consequence of the method used. Though there is extensive evidence of normal faulting and extension during the Jurassic and Cretaceous, the apparent throw on the faults cannot account for the 50–100 km extension required by the thermal model. If, however, the normal faults are listric faults, as the profile shown by Watson and Swanson [1] suggests, the extension may be concealed. The best estimate of the extension will probably come from seismic refraction, as it did in the Aegean. Because of the long history of extension in the North Sea, detailed modeling of the thermal structure of the underlying lithosphere will require several stretching events.

In basins even older than the North Sea, such as Michigan Basin, very little is known about the tectonic deformation of the underlying rocks [25]. What little information that is available is compatible with an extensional origin [5], and the size of the basin and its relationship to the Caledonian fold belt bears some resemblence to that of Aegean to the Alpide fold belt.

A thorough examination of large amounts of information, much of which is unpublished, is necessary to determine if the model works in detail. However, the preliminary discussion above suggests that such an examination is worth while.

4. Conclusions

The model discussed above produces a sedimentary basin by sudden stretching, followed by slow cooling of the lower part of the plate. It has a close resemblance to the thermal models which have been so useful in the oceanic regions. Such extension occurs in two geological environments: at rifted margins, such as the western European continental shelf and the Red Sea; and behind island arcs, such as the Aegean Sea and the Pannonian Basin [26]. The obvious objection to this suggestion is the amount of extension required: about a factor of two to produce a basin filled with 4.5 km of sediment. If the model is correct this extension must generally have been overlooked. Though plausible arguments are put forward as to how this could have happened, this is the least satisfactory part of the model. What is needed is a detailed study of the subsidence history, the heat flow, seismic refraction and reflection observations in reasonably large basin full of sediment. The basin should have been formed between about 10 and 50 My ago for the heat flow and subsidence history to give independent and accurate estimates of the extension and should be large enough to avoid flexural effects. The Pannonian Basin [26] seems suitable. An interesting feature of the model is the close relationship between the heat flux and the subsidence, both of which influence the chemistry of any hydrocarbons in the sediments.

It is important to point out that this model can only explain epeirogenic subsidence, not uplift above sea level. Structures such as the East African dome and the Colorado Plateau must have a different origin, and cannot be produced by passive upwelling of the asthenosphere into a broad zone between separating plates.

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