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The Strength of the Earth's Crust

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INTRODUCTION

Since its development during the mid-1960s the concept of plate tectonics has provided a framework in which to explain a wide range of observations in the Earth sciences. In this concept, the rigid outer layers of the Earth, or lithosphere, are viewed as comprising a number of major plates which move apart, converge, or slide past each other. Earthquakes, which themselves indicate plate motion, delineate the boundaries of the plates. Plate tectonics provides a description of the present seismic and tectonic activity at the Earth's surface. One of the objectives of geological and geophysical studies since the mid-1960s has been to apply the concept of plate tectonics to the geological past. As a result it is now possible to describe the geometry of major plate motions for about the past 200 m.y. of Earth history.

A fundamental premise of plate tectonics is that the boundaries of the plates are zones where differential movements between rigid plates occur. It is usually assumed in reconstructing past plate motions that the plates have behaved rigidly on very long geological time scales. But how rigid are the plates?

The principal evidence for the rigidity of the lithosphere on long time scales has come from studies of how it responds to surface loads such as oceanic volcanoes and river deltas. These loads are sufficiently long in duration (10^5-10^8 years) and of large enough mass (10^{21}-10^{22} gm) that they strain the lithosphere almost to the limits of its strength.

The purpose of this lecture is to briefly summarize some of the recent results of studies of the deformation of the lithosphere in the region of large geological loads and to discuss how these results may be used to determine information on the long-term rigidity of the Earth's outer layers. The discussion will be limited to oceanic regions since they have generally had a simpler thermal and mechanical evolution than the continents. Further, it is relatively easy in the oceans to distinguish features in the bathymetry, gravity field and geoid and, seismic data that can be directly attributed to lithospheric deformation. These data show that the lithosphere is sufficiently rigid that it can support large loads for long periods of geological time. However, its rigidity varies as a function of lithospheric age and load age. This lecture will examine the implications of these rigidity variations for better understanding and physical nature of the lithosphere and the tectonic evolution of the world's ocean basins.

DEVELOPMENT OF THE CONCEPT OF A STRONG RIGID OUTER LAYER OF THE EARTH

The concept of a strong lithosphere can be traced back to the development of the principle of isostasy, more than 130 years ago. G.B. Everest, while Surveyor-General to India, showed in 1847 that a discrepancy existed in triangulation surveys of the northern India plains between positions determined by geodetic and astronomic techniques. J.H. Pratt, as a result of an exhaustive study in 1854, suggested the discrepancy arose because the plumb-line that was used as a "reference" for the astronomic observations was locally deflected due to attraction of the Himalayan range to the north. But he was puzzled that the measured deflection was not as large as would be expected by the attraction of the mountains themselves. A few months later, in 1855, G.B. Airy suggested that the smaller deflection arose because the mass excess of the mountains was underlain by a competing mass deficiency at depth. Airy proposed that the mass deficiency took the form of a low density crustal "root" extending directly beneath the mountains. By analogy with timber blocks floating on water, he proposed that the Earth's crust was "floating" on an underlying "sea of lava." Airy pointed out though that while the Earth's crust was stong enough to support narrow loads, such as Mt. Schehallien in Scotland, it was unlikely to be able to support wide loads, such as the "Tablelands" of South Africa. Wide loads, in his view, were bounded by "fissures" or "breakages" and were supported more by the buoyancy of the crust than its strength.

Pratt disagreed with Airy, arguing in 1858 that the mass deficiency was caused by lateral variations in the density of the crust and upper mantle. In his opinion, individual "columns" of the Earth's outer layers beneath mountains were lower in average density than columns beneath the sea coast. Pratt's view was more in accord with the proponents of the contraction theory of the Earth who believed that mountains represented regions that have not cooled as much as adjacent, less elevated, regions.

The question of the strength of the Earth's crust may have become obscured in the disagreement between Airy and Pratt had it not been for the publication of a book entitled "The Physics of the Earth's Crust" by O. Fisher in 1881. Fisher, a strong opponent of the contraction theory, agreed with Airy that the Earth's crust rested on a weak substratum. In his view, mountains were the consequence of horizontal compression of a crust "with a certain degree of rigidity" that eventually would yield to compression by some form of plastic flow and crustal thickening. In the case of surface loads, Fisher suggested that the crust was sufficiently rigid that any "considerable" addition of load would cause a region to subside, while any removal of load would cause it to rise.

The geologist, C.E. Dutton, was most impressed with Fisher's book and wrote a complimentary review of it in 1882. In the review, Dutton defined "... the condition of the terrestrial surface which would follow from the flotation of the crust upon a liquid or highly plastic substratum ..." as isostasy. In 1888, Dutton stated that the principal geological observations supporting isostasy were sedimentation accompanied by subsidence and erosion followed by uplift, referring to these phenomenon as experiments "conducted by nature" from which it should be possible to "glean" information on the rigidity of the crust.

But these inferences of the role of crustal rigidity could not be proved by geological observations alone, although they were strongly suggestive of some role. The leading American and European geodesists at the time, however, required some type of crustal model in order to reduce the closure errors in geodetic surveys; a task of immense practical importance in mapping. They chose to parametize the Airy and Pratt models and found that these models successfully reduced their closure errors. But as parametized by J.H. Hayford and W. Bowie in 1912 and W.O. Heiskanen in 1924, these models ignored the strength of the crust.

Shortly after Dutton's paper, an American geologist, G.K. Gilbert, published the first detailed account of the raised shorelines of the former Lake Bonneville in the Basin and Range province of the western U.S. He showed that following removal of the water load the crust was uplifted by about 45 meters over an area of about $250 \times 600 km^2$. The amount of uplift was much less and the area of uplift larger than would be expected if the water was simply removed from a crust of no lateral strength. Gilbert interpreted this result as indicating that the crust resisted the uplift due to its lateral strength, modelling the uplift by an uparching of a relatively stiff elastic sheet.

But Gilbert's pioneering work on the strength of the crust was eclipsed by that of a younger contemporary, J. Barrell. In a series of papers published during 1914 and 1915, Barrell argued that the outer layers of the Earth were a good deal stronger than the advocates of "extreme" isostasy, such as Hayford and Bowie, would admit. Barrell, using theory developed earlier by G.H. Darwin, calculated the maximum stress difference implied by Hayford's isostatic gravity anomalies and showed that, depending on their size, they could be supported by the strength of the crust. He extended his analysis to topographic features arguing that narrow features, such as Hawaii and the Niger delta (full wavelength λ in the range 400-600 km), would mainly be supported by the strength of the crust while wide features, such as the Laurentide ice sheet ($\lambda \sim 2600$ km), would be largely supported by buoyancy of a weak underlying layer. In his early papers Barrell thought that the crust was strong enough that it would not yield by bending, even in the region of a large load such as the Nile delta. However, he argued in later papers that the crust would respond to most loads by bending and would in his words ". . . come to act to some extent as a bending plate." Barrell referred to this rigid outer layer of the Earth as the lithosphere, a term that is still in use today.

Barrell referred to the weak layer underlying the lithosphere as the asthenosphere. He believed that the asthenosphere was a zone of relatively low strength. However, it was not until a later study by W. Bowie, who showed that large river deltas were gravitationally compensated, that the existance of a weak substratum was specifically demonstrated.

The relative importance of the strength of the lithosphere—or as Gilbert posed the problem "isostasy versus rigidity"—continued to be vigorously debated following Barrell's death in 1919. The debate was focused in a special symposium held on Isostasy in 1922 sponsored by the Geological Society of America. One of the papers presented at the symposium was by the geologist, G.R. Putnam, who pointed out that gravity anomaly differences between closed spaced pairs of stations could be substantially reduced if the compensation of topography was not spread out locally as implied by local models of isostasy but was regionally distributed instead.

F.A. Vening Meinesz, a Dutch geophysicist, was influenced by Putnam's work and his preference for a regional form to isostasy. In 1931, he published a regional scheme of isostatic compensation which, in many ways, incorporated some of the earlier ideas of Gilbert and Barrell. He represented the Earth's topography as a load and calculated the corresponding deformation of the crust, assuming it could be modelled as a thin elastic plate overlying a weak substratum. The theory that he used was based on the work of H. Hertz who, several years earlier, had modelled the deformation of ice on ponds caused by skaters. By comparing calculated gravity anomalies based on this model to observed anomalies Vening Meinesz determined the "radius of regionality" for the crust, which he showed was a function of its flexural rigidity.

It was R. Gunn, however, who showed in a series of papers published between 1937 and 1947 that the model of an elastic plate could explain the deformation of the lithosphere in the vicinity of a wide range of geological features including passive continental margins, island arc-trench systems and mountain ranges. The differential equation for the deformation of an elastic plate overlying a weak fluid is

$$D \frac{d^4y}{dx^4} + (\rho_m - \rho_{infill})\, yg = 0 \qquad (1)$$

where x is the distance along the pre-deformation surface, y is the deformation, ρ_m is the average fluid density, ρ_{infill} is the average density of the material infilling the deformation and, g is average gravity. Imposing the boundary conditions that the deformation is small at great distances from an applied line load P and that it is continuous and symmetric beneath P we find that the solution of (1) is of the form

$$y = \frac{P \lambda e^{-\lambda x}}{2(\rho_m - \rho_{infill})g} (\mathrm{Cos}\, \lambda x + \mathrm{Sin}\, \lambda x) \qquad (2)$$

where λ is given by

$$\lambda = \left[\frac{(\rho_m - \rho_{infill})g}{4D} \right]^{1/4} \qquad (3)$$

and D is the flexural rigidity of the plate (Figure 1). By comparing the characteristic wavelength of flexural features, such as island arcs and trenches, to calculated deformation curves based on (2), Gunn argued that the lithosphere was strong enough that it could support the large stresses implied by the elastic plate model (>1 kbar) for long periods of geological time.

Curiously, it was another thirty years before plate tectonics, which also required a strong lithosphere, was developed and even then the arguments of Barrell, Vening Meinesz, and Gunn played little or no role in its development.

FLEXURE OF THE LITHOSPHERE

Since the development of plate tectonics there have been a number of detailed studies published on the bending, or flexure, of the lithosphere caused by surface loads. The general approach in these studies has been to assume the shape and mass of the load and compute flexure profiles based on an elastic plate model for different values of the flexural rigidity, D, and equivalent elastic thickness, T_e, of the lithosphere (e.g. Walcott, 1970). By comparing computed flexure profiles to observed topography, gravity and geoid, and seismic refraction data in the region of large loads it has been possible to estimate the value of T_e. One of the objectives of flexure studies in the oceans has been to determine the value of T_e and how it varies with tectonic setting and load age.

The volcanoes that rise up above the sea floor as seamounts and oceanic islands have proved particularly satisfactory loads for flexure studies. Oceanic volcanoes form relatively quickly on the lithosphere (as evidenced by the short time between eruptions) and occur in a variety of tectonic settings in the interior of the plates, far from the complexities of active plate boundaries. Further, the shape of a volcanic load can be relatively easily estimated from observed topography profiles since they typically build to heights of 3 to 4 km above "normal" ocean floor depths.

Figure 2 shows the flexure profile and the associated gravity anomaly that would be expected for a volcanic load the size of Oahu in the Hawaiian Islands. There is good overall agreement between observed gravity anomalies in the vicinity of Oahu and computed curves based on an elastic plate model with T_e = 30 km. The fit of the flexure profile to the observed topography and seismic data is not as good, however. The problem with the topography is that the Hawaiian islands are superimposed on a broad swell in the seafloor which in the region of Oahu is about 2000 km across and about 1 - 1.5 km above the normal depth of the seafloor. The swell, which may have resulted from reheating of the lithosphere by the Hawaiian hot-spot (Detrick and Crough,

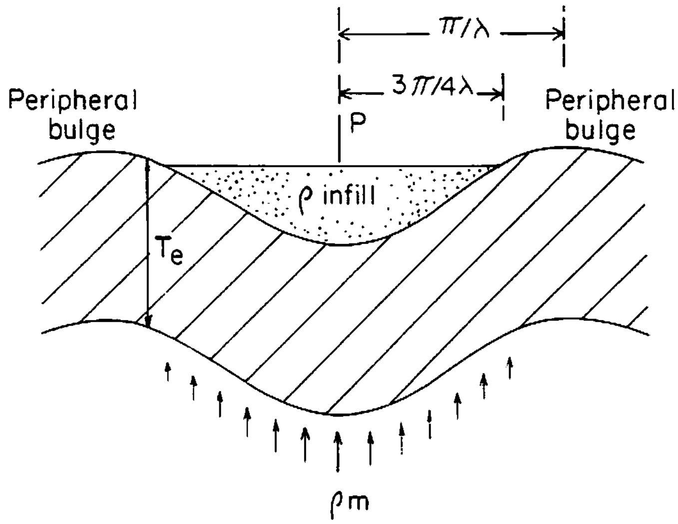

Figure 1. Schematic diagram of the flexure of a thin elastic plate overlying a weak fluid substratum due to a line load P. The equivalent thickness of the plate, T_e, is given by $(D12(1-\sigma^2)/E)^{1/3}$ where E is Young's Modulus and σ is Poisson's ratio. The surface load P is supported in part by the strength of the plate and in part by buoyancy of the weak underlying substratum.

1978) or from some form of convective flow pattern below the lithosphere (McKenzie et al., 1980), modifies the flexural bulges that developed during loading. The problem with the seismic data, on the other hand, is that is was obtained during the early 1960's and is generally unreliable because of poor navigation and drifting receivers. North of the ridge, seismic data gathered as part of Project "Moho" (Shor and Pollard, 1964) indicates that the depth to "Moho" dips gently from beneath the region of the Hawaiian arch to the moat. There is a close overall agreement between the depth to "Moho" and the flexure profile for $T_e = 30$ km (Figure 2). The seismic data beneath the ridge (Furumoto et al., 1968), however, is much too scattered (for example, the depth to the >7.8 km/sec reflector ranges from 6 to 14 km north of Oahu to about 20 km south of Oahu) to satisfactorily constrain the flexure profile.

There are, of course, uncertainties in the models that also complicate the comparison of observed and computed profiles. It was assumed in Figure 2, for example,

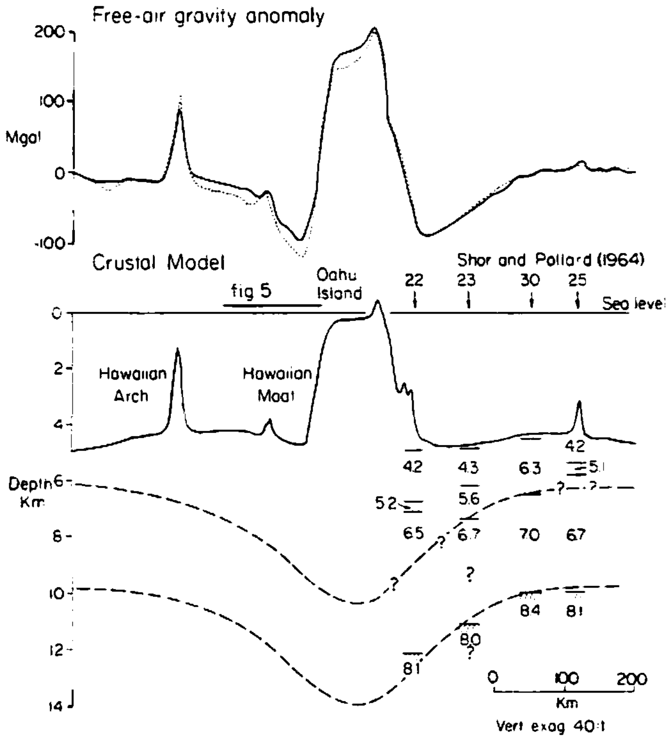

Figure 2. Comparison of the calculated gravity anomaly and flexure profile to observed free-air gravity anomaly and seismic refraction data in the region of Oahu island. The location of the gravity anomaly profile is given in Watts (1978) and is a composite of data obtained during R/V VEMA cruise 21 leg 5 and R/V ROBERT D. CONRAD cruise 12 leg 20. The dotted lines on the gravity profile and the dashed line on the seismic profile are based on a continuous plate model with $T_e = 30$ km. The seismic refraction stations, which have been projected normally onto the profile, are based on Shor and Pollard (1964). The heavy solid line indicates the approximate location on the profile of the seismic reflection profile shown in Figure 5.

that Oahu loaded a continuous elastic plate and that the density of the load (2.8 g/cm³) was equal to the density of the material infilling the depression. Yet other types of flexure models may also apply in the region of Oahu (Figure 3). Figure 4 compares the flexure profile and the associated gravity anomaly that would be expected for each model type. The largest differences occur between the continuous plate model with variable infill density and the fractured plate model. The continuous plate model, however, does not differ by more than ± 50 mgal from either of these two models. Thus, while the use of different flexural models will yield different values of T_e, they should not be significantly different.

In order to better constrain the value of T_e more detailed seismic refraction and reflection measurements will be required in the vicinity of large volcanic loads. During the summer of 1982 the Lamont-Doherty Geological Observatory and the Hawaii Institute of Geophysics conducted a multichannel seismic study in the region of Oahu and Molokai in the Hawaiian Islands. The experiment utilized two ships, one equipped with a 3.6 km multichannel seismic streamer and the other equipped with a large airgun array and explosive charges. A wide aperture single channel seismic reflection profile obtained during the experiment is shown in Figure 5. This figure shows that the flexural moat is associated with a number of prominent seismic reflectors which dip gently beneath the moat from the region of the flexural bulge towards the islands. These reflectors probably represent volcanoclastic material that infilled the flex-

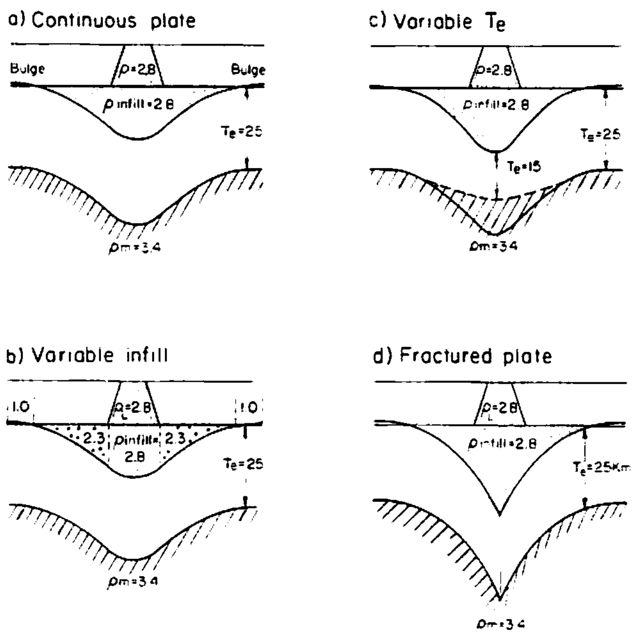

Figure 3. Simple models for flexure of the lithosphere caused by seamounts and oceanic island loads. (a) Continuous plate model. (b) Continuous plate model in which the density of the material infilling the flexure differs from the density of the load. (c) Continuous plate model in which T_e varies from 15 km beneath the load to 25 km in flanking regions. (d) Fractured plate model.

ural depression during loading. The two-way travel time to the deepest reflector is about 1 sec below the seafloor which corresponds to a depth of about 2.5 km below the seafloor (assuming a mean velocity of the infill of 5.0 km/sec). However, all the seismic data obtained during this experiment will need to be processed before this data can be satisfactorily used as a constraint for the flexure model.

A number of studies have been carried out of other types of loads on oceanic lithosphere including mid-ocean ridge crest topography, island arc—deep-sea trench systems, and river deltas. Of these load types particular attention has been focused on river deltas. Like seamounts and oceanic islands, river deltas form relatively quickly on the plates and occur in a variety of tectonic settings. The main disadvantage of these loads, however, compared to seamounts and oceanic islands is that it is not as easy to estimate the shape of the load, since the "normal" depth of basement is not generally known in the region of a river delta.

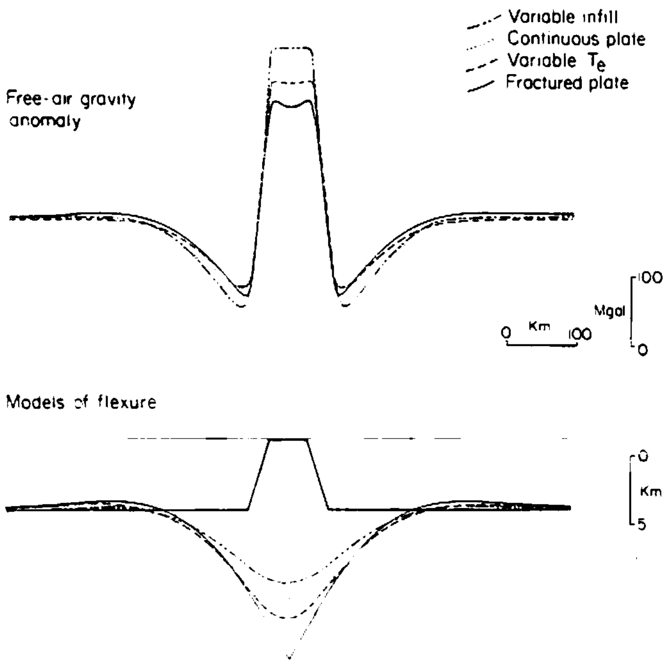

Figure 4. Calculated gravity anomaly and flexure profiles for each of the flexure models in Figure 3. The largest amplitude gravity anomalies and smallest amplitude flexure occurs for model (b) in which the density of infill material is assumed lower than that of the load. The smallest amplitude anomalies and largest flexure occurs for model (d) in which the plate is fractured directly beneath the load. The maximum differences between the models occur in the region of the seamount load. Differences are small (<20 mgal; <1.5 km) in regions flanking the seamount and it would be particularly difficult to distinguish between the models from observations din these regions. A continuous plate model (dotted line) appears representative of the model profiles and its associated gravity anomaly does not differ from other models by more than 50 mgal over the seamount and 25 mgal over flanking regions.

Figure 6 shows the flexure profile and the gravity anomaly expected for a sediment load the size of the Amazon river delta, off Brazil. The computed curves, like those for Oahu, assume an elastic plate model with T_e = 30 km. The dotted line on the crustal model indicates the load shape that was assumed for the delta. The region above the dotted line represents the load while the region below the line represents the flexural depression. Figure 6 shows there is good agreement between the computed and observed gravity anomaly except over the inner part of the shelf. The fit of the flexure profile to the seismic data is not as good, although the data that is available (Houtz et al., 1977) indicates the depth to the top of oceanic basement increases steeply beneath the continental rise and slope, in general agreement with the predictions of the flexure model.

The estimates of T_e from all oceanic flexure studies are plotted in Figure 7 against the age of the lithosphere at the time of loading. The estimates are based on different types of loads and were obtained by comparing

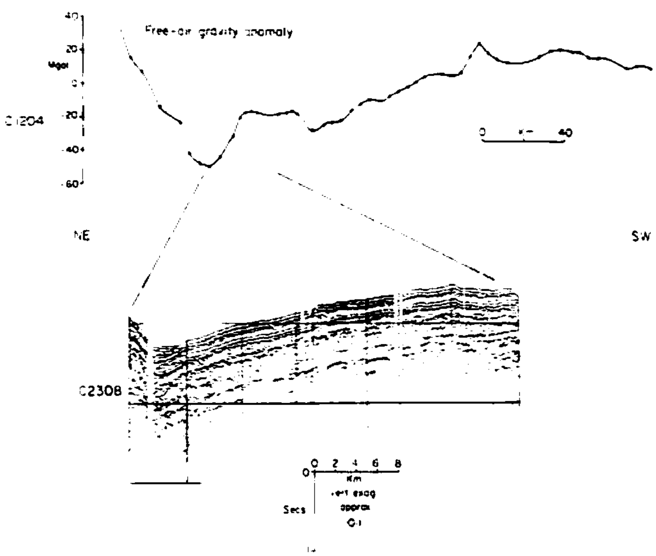

Figure 5. Single-channel wide aperture seismic reflection profile in the region of the gravity anomaly low flanking the south side of Oahu island. The profile was obtained during summer 1982 on R/V *Robert D. Conrad* cruise 23 leg 8 during a joint Lamont-Doherty Geological Observatory and Hawaii Institute of Geophysics seismic study of lithospheric flexure along the Hawaiian-Emperor seamount chain. The profile was obtained with a 3.5 km long 48 channel seismic streamer (only the monitor record from one channel in the mid-section of the streamer is shown) and a large volume (2000 cu.in.) air gun array. Multichannel seismic reflection profiles elsewhere in the Pacific usually show reflections from the "Moho" at about 2 seconds two-way travel time below the seafloor. The strong seismic reflectors at depths of up to 1 second below the seafloor in the flexural depression represents stratification of material that infills the flexural depression.

observations of topography, gravity and geoid, seismic refraction or uplift/subsidence data to calculated profiles based on an elastic plate model. The loads included seamounts and oceanic islands (circles), river deltas (diamonds), deep sea trench—outer rise (triangles) and mid-ocean ridge crest topography (squares). The age of the lithosphere at the time of loading was estimated by subtracting the age of the load from the age of the adjacent oceanic lithosphere. For example, in the case of Oahu, the age of the oceanic lithosphere in the region of the island is approximately 80 m.y., the age of the load is about 3 m.y., and the age of the lithosphere at the time of loading is about 77 m.y.

There are two main conclusions that may be drawn from the data plotted in Figure 7. First, T_e increases with the age of the lithosphere at the time of loading. The simplest interpretation of this result is that as the lithosphere cools it becomes progressively more rigid in the way it responds to surface loads (Watts, 1978). Loads formed on young lithosphere, such as the Walvis ridge guyots, are associated with relatively low values of T_e while loads formed on old lithosphere, such as the Hawaiian Islands, are associated with relatively high values. There is good general agreement between T_e and the depth to the 300 °C to 600 °C ocean isotherms, based on the cooling plate model, suggesting the rigidity is determined by the thermal structure of the lithosphere. Second, T_e is 2 to 3 times smaller than the seismic thickness of the lithosphere based on short-term seismic wave periods. Thus, since the loads plotted in Figure 7 range from 0.5 m.y. to 58 m.y., there must be a rapid relaxation of the thickness of the lithosphere that supports a load, from its short-term (~150 secs) or seismic thickness to the long-term mechanical thickness. Subsequent relaxation does not appear to occur, although it may occur on very long time scales.

These conclusions are consistent with cooling plate models (e.g., Parsons and Sclater, 1977) which predict that the thermal thickness of the lithosphere increases with time. The material that is added to the base of the lithosphere as it cools is not altered by the original load and would not be expected to alter the value of T_e that is acquired soon after loading. However, for subsequent loads, such as renewed volcanism, the lithosphere would deform with a T_e value that depends on the new depth to the 300 °C to 600 °C oceanic isotherms.

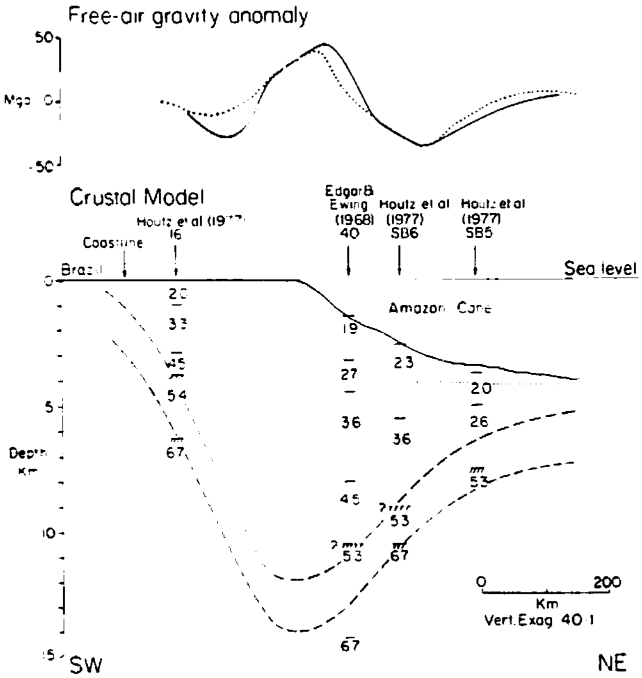

Figure 6. Comparison of the calculated gravity anomaly and flexure profile to observed free-air gravity anomaly and seismic refraction data in the region of the Amazon Cone. The location of the gravity anomaly profile is given in Cochran (1973). The dotted lines on the gravity profile and the dashed lines on the crustal model are based on a continuous plate model with T_e = 30 km. The dotted line on the crustal model defines the load that was used to compute the flexure profile.

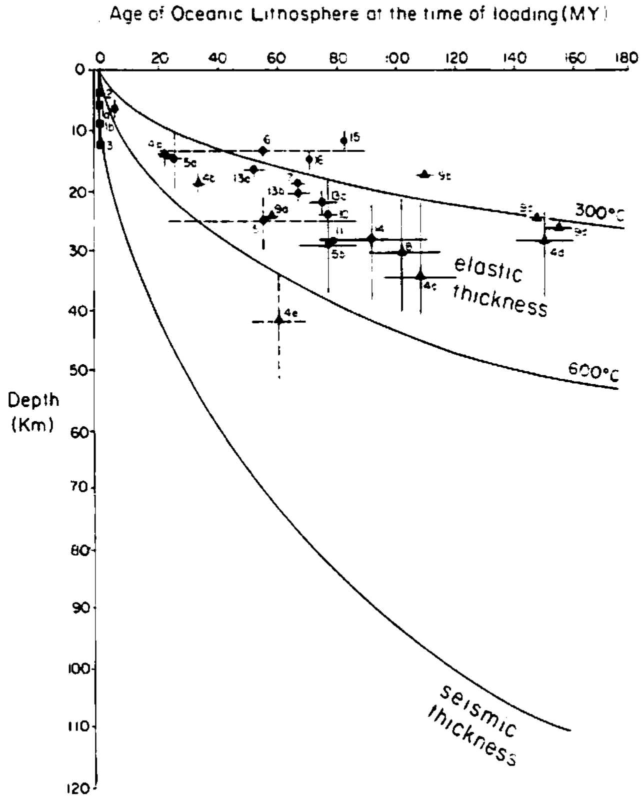

Figure 7. Plot of elastic thickness, T_e, against age of oceanic lithosphere at the time of loading. The values of T_e (1-14) are based on different loads on the lithosphere summarized in Watts (1982). The plot includes two additional estimates (15,16) by Lambeck (1981a, 1981b) from the Southern Cook and Society Islands. The age of the lithosphere at the time of loading for each feature was obtained by subtracting the age of the feature from the inferred age of the underlying sea-floor. The solid lines that envelope the T_e estimates are based on a cooling plate model for oceanic lithosphere (Parsons and Sclater, 1977). The seismic thickness of the lithosphere is based on surface wave and long-range seismic refraction studies.

The results of oceanic flexure studies have been expressed here in terms of the parameter T_e which refers to the stiffness of a purely elastic plate. This is not to imply that the materials of the lithosphere would behave purely elastically in response to loads. Figure 7 suggests, in fact, that T_e changes rapidly with time following loading, approaching an asymptotic value that depends on the age of the lithosphere. The lithosphere only appears elastic, therefore, after a certain amount of time has elapsed (about 10^6 years) and even then, the thickness of the elastic layer is largely determined by its age.

The oceanic lithosphere is apparently rigid enough that it can support large geological loads for long periods of time. The rigidity of the lithosphere is determined largely by its age, however. Oceanic lithosphere appears as a strong structure for old ages and a weak structure for young ages. There is an absence of faulting in the flexural depression flanking large volcanic loads (e.g., Figure 5), although faulting has been reported in the seaward wall of deep-sea trenches (e.g., Ludwig et al., 1966). By itself the value of T_e does not therefore provide a measure of the ultimate strength of the lithosphere. However, the relationship of T_e to plate age provides important constraints on the physical nature of the lithosphere.

THE PHYSICAL NATURE OF THE LITHOSPHERE

The widespread occurrence of isostasy indicates that overall, the Earth's crust is in a state of hydrostatic equilibrium. This implies that on the scale of continental masses the crust "floats" on the underlying mantle, in a similar way as envisaged earlier by Airy and Fisher. But the pressures under which an individual column of the crust is in equilibrium are determined by the stresses that act on it due to its weight and the buoyancy of the underlying material. Surface loading would upset the equilibrium, causing the column to move until equilibrium is restored. The effect of rigidity is to resist these movements and "re-direct" how equilibrium is restored. Thus, departures from Airy isostasy, such as flexure, provide information on the rigidity of the crust.

It was pointed out earlier that on loading, the thickness of oceanic lithosphere supporting a load must relax rapidly from its short-term to its long-term thickness. This implies that some form of stress relaxation occurs in the lithosphere soon after loading. Thus, the oceanic lithosphere shows characteristics of both elastic and viscous materials. Rather than use a purely elastic model, therefore, it may seem more appropriate to consider a model that combines elastic and viscous behavior, such as a viscoelastic (Maxwell) model. In an elastic model, T_e determines the shape of the flexure profile and depends only on the age of the lithosphere at the time of loading. The viscoelastic (Maxwell) model, on the other hand, is described by an initial, or zero time, elastic thickness, T_o, and a Maxwell relaxation time, τ. The shape of the flexure profile is given by T_o and τ and depends on the age of the load.

In order to evaluate the question of whether the oceanic lithosphere is elastic or viscoelastic on long time scales let us first consider its response to a volcanic load approximately the size of Oahu. For an elastic model the lithosphere would respond with a rigidity that depended on its age. In the case of a viscoelastic model, however, the lithosphere would respond as a relatively thick, rigid, plate for small load ages (compared to the Maxwell time) and as a thin, weak, plate for large load ages. Thus, the principal means to distinguish between the two types of models is to determine whether T_e depends mainly on the age of the lithosphere at the time of loading (elastic model) or the age of the load (viscoelastic model).

The principal facts relating to the age and the elastic thickness, T_e, for four volcanic loads are summarized in Table 1. The loads have been grouped in the table into the Hawaiian-Emperor seamount chain in the central and northwest Pacific and the Southern Cook and Society Islands in the southwest Pacific. The oldest sea-

TABLE 1

	Age of Load t m.y.B.P.	Age of Lithosphere t_L m.y.B.P.	Age of Lithosphere at the time of loading $(t_L - t)$ m.y.	T_e* km	$\tau = 10^6$ yrs. T_o km	Reference for T_e, t and t_L
Emperor Seamounts North of 40°N	52-58	~80	22-28	10-20	45-86	Watts (1978)
Hawaiian Ridge	3-18	80-105	72-87	17-37	39-89	Watts (1978)
Rarotonga (Cooks)	2-3	~85	82-83	10-13	16-20	Lambeck (1981a)
Tahiti (Society)	0.5-1.5	~70	68-69	21	26	Cazenave et al. (1980)

*Assumes E = Youngs Modulus = 10^{12} dyne/cm^2 and σ = Poisson's ratio = 0.25.

mounts in each group correlate with the lowest values of T_e, an observation that apparently supports the viscoelastic model. By specifying the load shape, T_e can be used to estimate the equivalent T_0 and τ, provided one of them is known. Let us consider that the parameter τ is known for oceanic lithosphere and has a value of 10^6 years. By assuming τ, the value of T_0 can then be estimated for each load in Table 1. The age of the Emperor Seamounts is significantly greater than the assumed τ (50 to 60 times τ) so the lithosphere in the vicinity of this load would be expected to show the greatest relaxation and the corresponding initial thickness, T_0, would be 4 to 8 times T_e. The age of Tahiti (Society Islands), however, is of the same order as τ so the lithosphere would be expected to show the least relaxation and the corresponding T_0 would only be 1 to 1.5 times T_e. This suggests large variations in the value of T_0 at these loads. Such large differences in T_0 would not be expected, however, since the age of the lithosphere on which these loads formed is not significantly different. The T_e for these loads cannot therefore be easily explained by a viscoelastic model with a uniform value of $\tau = 10^6$ years. In fact, in order to fit the values of T_e for Emperor Seamounts, a relatively high τ would be required (10^5-10^6 years) while for Tahiti a much smaller τ would be needed ($\sim 10^3$ years).

These arguments imply that a viscoelastic (Maxwell) model does not adequately describe the response of oceanic lithosphere to long-term loads, unless a value of τ is used that depends on the load age. Instead, the low values of T_e for the northern Emperor seamounts can be explained by an elastic model, since these features formed on relatively young lithosphere. However, the low value for Rarotonga in the Southern Cook Islands cannot be as easily explained since they apparently formed on relatively old lithosphere. One explanation, which is supported by geological evidence (Wood and Hay, 1970), is that the volcanic history of the Southern Cook Islands group as a whole began earlier, in Paleocene to Oligocene times. Thus, the low value of T_e for Rarotonga may represent an *average* response of the lithosphere to a Paleocene/Oligocene loading event on young sea-floor and a Pliocene event on older sea-floor.

The actual mechanical properties of the lithosphere are likely, of course, to be more complex than would be predicted by a simple elastic model. The strength of a purely elastic plate is theoretically unlimited whereas the actual materials of the lithosphere are likely to have a finite strength.

The question that remains, therefore, is whether the results of flexure studies can be explained in terms of current knowledge based on extrapolations of data from experimental rock mechanics. These data suggest the strength of the lithosphere is limited by brittle failure and ductile flow. Rock friction experiments (e.g., Byerlee, 1968) indicate that the shear stress required to overcome static friction between rock surfaces increases as the normal stress across the surface increases. This implies that the strength of rocks increases with depth. The strength of rocks cannot continue to increase with depth, however, since as pressure and temperature increases, rocks become ductile and creep (Goetze, 1978). One of these types of creep, steady state creep, increases greatly with pressure but is rapidly accelerated by temperature. This implies a decrease in strength of the lithosphere with depth.

The experimental data from rock mechanics have been combined into a "yield stress envelope" model for the lithosphere (Goetze and Evans, 1979). In this model, the upper slope of the envelope defines the brittle field while the lower slope defines the field for ductile flow. The ductile field, unfortunately, can only be studied in the laboratory for a small range of strain rates from 10^{-4} to 10^{-6}/sec. But Goetze (1978) predicted the ductile field for the strain rates of 10^{-14} to 10^{-16}/sec, more appropriate to geological processes. For example, for an applied stress of 1 to 10 Kbar, he predicted that ductile behavior would occur at temperatures as low as 500 to 800 °C for a strain rate of 10^{-14}/sec.

A "yield stress envelope" model appropriate for 80 m.y. oceanic lithosphere is shown in Figure 8. The upper slope defines the brittle deformation field while the lower slope defines the ductile field. The shaded region is an estimate of the stress profile required to explain the bending moments that developed beneath Oahu during loading. Because there is limited seismic refraction data available beneath Oahu it is not possible to estimate these moments directly from seismic data. Instead, the bending moments were estimated from the elastic plate model for $T_e = 20$ and $T_e = 50$ km. Figure 8 shows that the thickness of the lithosphere that supports the Oahu load is limited by brittle deformation and ductile flow. However, the lithosphere has sufficient strength at depth that it can support the load without failure. In particular, the Oahu load is supported by an elastic "lid" which can maintain stresses of up to 2 kbar without failure. Figure 8 shows the "lid" thickness is similar to the elastic thickness, T_e, for 80 m.y. lithosphere. Further, since the "lid" thickness is strongly controlled by temperature, T_e would also be expected to be a strong function of temperature. Thus, in the case of the Oahu load, the data from experimental rock mechanics are in close agreement with the predictions of the elastic plate model.

The elastic plate model emerges from these discussions as the most useful "first-order" model describing the response of the lithosphere to surface loads. The phenomenon of loading is an important geological process which has occurred frequently during the history of the Earth. Thus, these inferences of lithospheric mechanics have important implications for better understanding the tectonic evolution of the world's ocean basins and their margins.

TECTONIC IMPLICATIONS

Oceanic flexure studies suggest that the response of the lithosphere to long-term loads can be satisfactorily modelled as a simple elastic plate of characteristic thickness T_e. These studies are mainly based on geological and geophysical data in the region of loads, such as the Hawaiian Islands and Amazon cone, which are of known tectonic setting and where loading is the dominant geological process. Thus, flexure studies have important implications for geological features, such as isolated seamounts, which are of largely unknown tectonic setting, and for features, such as passive continental margins, where processes of stretching are occurring in addition to loading.

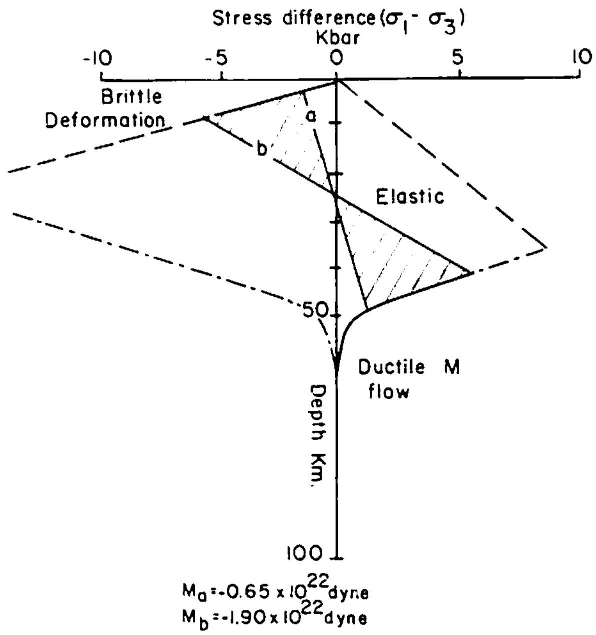

Figure 8. Yield stress envelope (Goetze and Evans, 1979) for 80 m.y. oceanic lithosphere. The upper slope is based on Byerlee's law in which the shear stress τ needed to break static friction is given by $0.6\sigma_n$ where σ_n is the normal component of stress. The slopes for tension (0.22 bars/km) and compression (0.66 bars/km) were estimated using a Mohr's circle with an average density of the lithosphere of 3.3 g/cm^3, an angle between the normal to the shear and direction of maximum principal stress of 30° and, g = 981 cm/sec^2. The lower slope is based on a Power law in which strain rate, $\dot{\varepsilon}$, is given by $70\ (\sigma_1 - \sigma_3)^3 e^{-Q/RT}$ where Q is the creep activation energy, R is the Gas constant and T is temperature. The slopes for tension and compression where obtained with Q = 122 Kcal/mol and $\dot{\varepsilon} = 10^{-16}$/sec. The shaded region indicates the stress-differences required to support the load of Oahu (Watts et al., 1980b) for two cases of T_e = 20 km (a) and T_e = 50 km (b). The positive x axis indicates tension and the negative axis compression. Note that the Oahu island load is not large enough to cause the entire plate to fail. The load is apparently supported instead, by a 30-45 km thick elastic "lid." The inferred lid thickness, which is limited by brittle deformation in the crust and ductile flow in the underlying mantle, is only slightly smaller (17-37 km) than the T_e inferred from flexure studies.

It is useful first though to better quantify the role of flexure at geological features. Earlier it was argued that T_e is controlled by the age and hence thermal structure of the lithosphere at the time of loading. The lithosphere appears as a weak structure for loads that form on or near a mid-ocean ridge crest and as a strong structure for loads that form off-ridge. However, the actual amplitude and wavelength of the flexure profile depends not only on T_e but also on load shape.

Let us consider, for example, the case of a harmonic load applied to the surface of an elastic plate overlying a weak fluid substratum. If the load is of the form h Cos (kx) where h is the maximum load height, x the horizontal distance and, k is the wavenumber ($k = 2\pi/\lambda$, where λ is the equivalent load wavelength) then its mass is given by,

$$P(x) = (\rho_L - \rho_W)\, g h \operatorname{Cos}(kx) \qquad (4)$$

where ρ_L is the average load density and ρ_W is the average density of material displaced by the load. The response of the lithosphere to the load is also be harmonic in the form (e.g., Officer, 1979) and is given by,

$$y(x) = \frac{(\rho_L - \rho_W)\, h \operatorname{Cos}(kx)}{(\rho_m - \rho_{infill})} \left[\frac{D k^4}{(\rho_m - \rho_{infill}) g} + 1 \right]^{-1} \qquad (5)$$

$$= \frac{(\rho_L - \rho_W)\, h \operatorname{Cos}(kx)}{(\rho_m - \rho_{infill})}\, \Phi \qquad (6)$$

Equation (5) shows that for the case of long wavelength loads ($k \to 0$, $\lambda \to \infty$), $\Phi \to 1$, and

$$y(x) \to \frac{(\rho_L - \rho_W)\, h}{(\rho_m - \rho_{infill})} \qquad (7)$$

If $\rho_L = \rho_{infill} = \rho_{crust}$ (which is the case for a crust of uniform density equation (7) gives the response of an Airy-type crust or very weak plate to a load. Thus, the lithosphere appears as a weak structure for very wide loads. For the case of short-wavelength loads ($k \to \infty$, $\lambda \to 0$), $\Phi \to 0$ and, $y(x) \to 0$ which gives the response of an infinitely rigid plate to a load. Thus, the lithosphere appears as a strong structure for narrow loads. For infinite values of k the plate yields by flexure. Hence, the function Φ completely determines the response of the lithosphere for particular values of k and T_e (and hence D).

Figure 9 shows a plot of the response function Φ against wavenumber for values of T_e that correspond to relatively young (T_e = 5 km) and old lithosphere (T_e = 25 km). Since most loads that are emplaced on oceanic lithosphere can be characterized by these values of T_e Figure 9 illustrates the role of flexure in the oceans for different load wavelengths. Apparently, flexure in the oceans is most important for load wavelengths in the range $50 < \lambda < 1000$ km. Loads of greater wavelength would be almost entirely supported by the buoyancy of the weak fluid underlying the lithosphere, while shorter wavelength loads would mainly be supported by the strength of the lithosphere. The oceanic lithosphere therefore acts to some extent as a "low-pass filter" in the way it responds to loads, its strength suppressing the short-wavelength Airy-type response but passing the long-wavelength flexural-type response.

The typical wavelength range represented by seamounts and sedimentary basin loads are also shown in Figure 8. The wavelength range for seamounts includes volcanic features such as isolated seamounts and some aseismic ridges, while the range for sedimentary basins includes sedimentary features such as river deltas and wide continental margin basins. Figure 8 suggests that, depending on their tectonic setting, lithospheric flexure plays an important role at these features. Because of their large horizontal extent, however, flexure will be less important at oceanic plateaus and the well sedimented abyssal plains. Similarly, because of their small horizontal extent, flexure will be less important at mid-oceanic ridge crest generated topography and narrow continental shelf basins.

These considerations imply that the oceanic flexure studies in Figure 7 have important additional implications for seamounts and sedimentary basins. These features occur in a sufficiently wide range of tectonic settings and horizontal spatial scales for flexure to provide a major control on their structure and evolution.

Seamounts are widely distributed in the ocean basins, occurring in both on or near ridge crest and off-ridge tectonic settings. Figure 7 suggests that seamounts formed on or near a mid-ocean ridge crest should be associated with relatively low values of T_e while seamounts formed off-ridge should correlate with relatively high values. Thus, by determining the value of T_e at a seamount of unknown tectonic setting it should be possible to estimate whether it formed on or near a ridge crest or off-ridge.

It was pointed out earlier that the gravity anomaly is a sensitive indicator of the value of T_e at a surface load. Figure 10 compares the gravity anomaly and the equivalent undulation of the geoid that would be expected for two identical sized seamounts; one formed on or near a mid-ocean ridge crest (Seamount A) and one formed off-ridge (Seamount B). The amplitude and wavelength of the gravity anomalies are significantly different for the two seamounts. Thus, it should be possible to use the Earth's gravity field to estimate the tectonic setting of seamounts of unknown origin.

A number of studies have now been carried out that have used the gravity field to estimate the tectonic setting of oceanic features. In the Pacific Ocean, Watts et al. (1980a) have shown that the Hess rise, Line islands ridge, Necker ridge, Mid-Pacific mountains, and Manihiki plateau all formed on or near a mid-ocean ridge

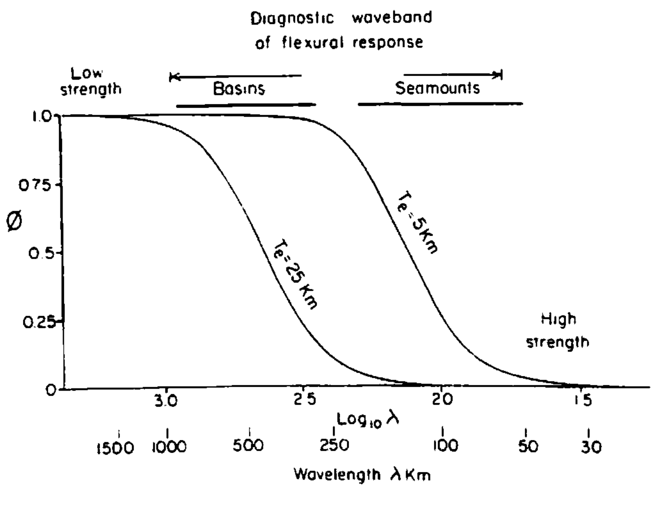

Figure 9. Plot of the response function Φ of an elastic plate overlying a weak fluid substratum for $5 < K < 240$ km^{-1}, corresponding to full wavelengths in the range 30 to 1500 km. The figure shows the diagnostic waveband (and equivalent wavelengths) of the flexural response of an elastic plate for values of T_e of 5 and 25 km. The wavelength band is in the range 50 to 250 km for $T_e = 5$ km and 150 to 1000 km for $T_e = 25$ km. Thus, flexure is likely to be important at seamounts and sedimentary basins which have "characteristic" wavelengths of 50 to 200 km and 200 to 1000 km respectively.

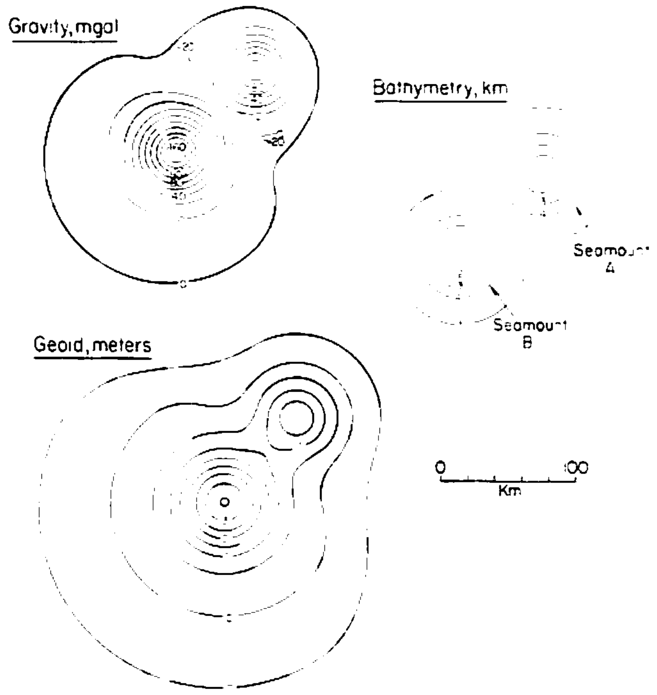

Figure 10. Calculated gravity and geoid anomaly for two isolated seamounts; one formed on relatively young ocean floor (T_e = 5 km) and one formed on old ocean floor (T_e = 25 km) (based on Watts and Ribe, in prep). The model calculations predict that seamounts formed on young lithosphere should be associated with relatively short-wavelength, small amplitude, gravity (~100 mgal) and geoid (~2 meters) anomalies while seamounts formed on old lithosphere should be associated with relatively long-wavelength, large amplitude, gravity (~160 mgal) and geoid anomalies (~4 meters).

crest. These features are mainly of volcanic origin and are located on 90 to 120 m.y. oceanic lithosphere suggesting there was an extensive outpouring of volcanic material onto the accreting Pacific plate boundary during Late Cretaceous times. The Magellan seamounts and part of the Marcus-Wake guyots, on the other hand, formed off-ridge, some distance from a mid-ocean ridge crest. Although these features are also of volcanic origin and formed on ~140 m.y. lithosphere, their age is still uncertain. They may be similar in age to the Late Cretaceous volcanic event on the accreting Pacific plate boundary. Alternately, they may represent a much younger event. In the Atlantic Ocean, Detrick and Watts (1979) showed that the Walvis Ridge formed on or near a ridge crest while Roberts and Bott (in press) have shown that the Wyville-Thompson Ridge formed off-ridge on 30-40 m.y. lithosphere.

Like seamounts and oceanic islands, sedimentary basins occur in a variety of tectonic settings in the ocean basins. Most basins can be classified as occurring either in a rift or orogenic setting. The main factors controlling the development of rift basins are thermal contraction following heating and thinning of the lithosphere at the time of rifting and, sedimentary loading. Figure 7 suggests that sediments formed early in basin evolution, when the lithosphere is relatively hot and weak, will be associated with low values of T_e while sediments which form later in margin history, when the lithosphere is relatively cold and strong, will be associated with high values of T_e. Thus, following rifting, sediments would be expected to progressively onlap the basement as it cools and increases its flexural strength with time.

An example of the stratigraphy that would be expected at the edge of a rifted sedimentary basin is shown in Figure 11. In the model it was assumed that the heating and thinning occurred by crustal and lithospheric extension at the time of rifting (McKenzie, 1978) and sedimentary loading occurred by flexure of a progressively more rigid basement following rifting. Initially, there is a rapid onlap of sediments onto the basement, due to the rapid transition that is assumed from extension to cooling, followed by a downward shift in onlap due to lateral heat conduction from the hot stretched region to the cool unstretched region. After about 16 m.y. flexure overcomes the effects of lateral heat flow and sediments progressively onlap the cooling basement.

Figure 11 suggests that coastal onlap should be a characteristic tectonic-stratigraphic feature of sedimentary sequences in rifted continental margin basins. This conclusion is of importance because measurements of coastal onlap from multichannel seismic reflection profiles of passive margins were the principle means by which Vail et al (1977) estimated sea-level rise through geological time. These workers took into account a regional tilting of the basement at the edge of a basin, but they did not account for flexure of a basement that increases its rigidity with age. Thus, the curves of relative changes of coastal onlap determined by Vail et al. (1977) may include tectonic as well as eustatic effects.

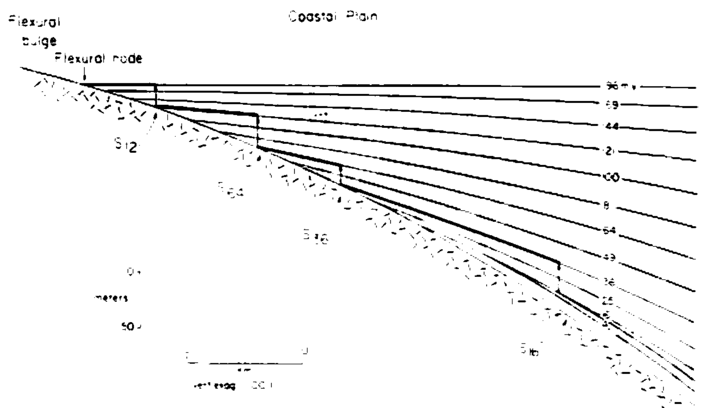

Figure 11. Calculated stratigraphy at the edge of a sedimentary basin formed by heating and thinning of the lithosphere at the time of rifting (based on Watts, 1982). The solid lines illustrate the procedure that would have been used by Vail et al. (1977) in the absence of sedimentary facies information to estimate sea-level changes during geological time. The model suggests that the progressive onlap of strata onto the basement at the edge of a rifted basin could be caused by lithospheric flexure and does not require sea-level changes to produce it.

The question of the nature of the control on stratigraphic sequences, remains an important outstanding problem. It was argued by Vail et al (1977) that the main control is eustatic because many patterns of onlap and offflap are widespread. However, as Figure 10 shows, tectonics in the form of flexure can give rise to similar patterns of onlap and offlap. Further, flexure explains the characteristic wedge shape, or "steer's head" (Dewey, 1981), observed at the edges of many basins. Following formation of a rift basin T_e increases with time so we may expect flexure to provide a characteristic signature of rhythm to basin stratigraphy. Thus, since many widely separated basins rifted at similar times, the tectonic control on stratigraphy could be widespread and difficult to separate from worldwide eustatic effects.

SUMMARY

The study of the deformation of the lithosphere caused by surface loads has provided important information on the long-term mechanical properties of the lithosphere. The model that best describes geological and geophysical observations in the vicinity of large loads is an elastic plate model, similar to one suggest earlier by Gunn and Vening Meinesz. In this model, the lithosphere responds to long-term loads ($>10^5$ – 10^6 years) in a similar manner as a thin elastic plate overlying a weak fluid substratum. Gunn and Vening Meinesz showed that the form of the deformation, or flexure profile, is controlled by the flexural rigidity and the equivalent elastic thickness, T_e, of the plate.

The principal contribution of oceanic flexure studies during the past decade has been to estimate the value of

T_e for a number of different types of loads on the lithosphere. T_e has been shown to decrease rapidly during loading, approaching an asymptotic value that depends on the age of the lithosphere. The oceanic lithosphere appears as a relatively weak structure for young ages and a relatively strong structure for old ages. These results have important implications for the physical nature of the lithosphere and the tectonic evolution of the world's ocean basins.

Flexure studies represent a new quantitative approach to geology in which an attempt is made to explain tectonics in terms of the mechanical properties of the lithosphere. This approach offers much promise during the next decade to better understand geological features, such as sedimentary basins and mountain ranges, where processes of stretching and compression are occurring in addition to loading, as well as some of the fine details of stratigraphy and structural geology.

But in order to understand the mechanical properties of the lithosphere it will be necessary in the future to obtain better constraints on the physical properties of the crust and upper mantle in the vicinity of large loads. The oceans appear to be the best environment in which to carry out such a program because they have had a simpler physical and chemical evolution than the continents. Further, the oceans include a number of examples of well preserved surface loads and it is relatively easy to distinguish their flexural effects in observed topography, gravity and geoid, and seismic refraction and reflection profile data. The program should focus initially on short-duration loads, such as volcanoes and river deltas. Loads on young (e.g., Emperor Seamounts) and old (e.g., Amazon Cone) lithosphere should be studied. Later the program should focus on regions of the lithosphere, such as the East Coast U.S. and Gorringe Bank, where loading is occurring in addition to other geological processes such as stretching and compression. Seismic reflection and refraction profiling using wide aperture multichannel arrays and large sound sources to define the detailed velocity structure and depth to Moho in loaded regions, in combination with deep crustal drilling to determine the nature of epeirogenic movements in these regions, provide the most promising techniques with which to study these loads during the next decade.

It may be pointed out that plate tectonics was developed largely from geological and geophysical data collected in the oceans. Yet plate tectonics continues to have a major impact on our understanding of the continents. There is no reason to suppose that a vigorous program of lithospheric mechanics in the oceans during the next decade should not have an equally great impact.

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