NOTES

Subsidence of a guyot west of Flores, Azores

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The analysis of mechanisms of subsidence of seamounts is often limited by inadequate information on the age of the seamount at the time it was first eroded to sea level. The ages of carbonates deposited on the eroded surface and the ages of the basalts therefore provide valuable observations for evaluating subsidence models. Such ages were recently reported for a small seamount in the Azores. It was found that the resulting subsidence was 100–150 m greater than could be attributed to thermal contraction of the ocean floor. This discrepancy can be attributed to stress relaxation within the underlying lithosphere. Viscoelastic models predict a subsidence that is a function of the effective relaxation time τ and the effective flexural rigidity D. Consistent parameters for 5–10 Ma old lithosphere are $\tau \simeq 0.5-1.0$ Ma and $D \simeq 10^{21}$ N·m.

L'analyse des mécanismes de subsidence des guyots est souvent limitée par le manque de renseignement sur l'âge du guyot au moment de son érosion au niveau de la mer. Les âges des carbonates sédimentés sur la surface érodée combinés avec les âges des basaltes peuvent donc fournir des données utiles pour évaluer les modèles de subsidence. De tels âges furent rapportés récemment pour un petit guyot dans les Açores. Les résultats indiquent que la subsidence dépassait de 100–150 m celle qui serait attribuée à une contraction thermique du fond marin. Cette différence peut s'expliquer par une relaxation des contraintes au sein de la lithosphère sous-jacente. Les modèles de viscoélasticité prédisent une subsidence qui est en fonction du temps de relaxation effective τ et de la rigidité de flexure effective D. Les paramètres de la consistance d'une lithosphère formée il y a 5–10 Ma sont $\tau \simeq 0{,}5-1{,}0$ Ma et $D \simeq 10^{21}$ N·m.

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Introduction

Ryall *et al.* (1983) have investigated in considerable detail a small seamount located in the North Atlantic some 50 km west of Flores, the most westward island of the Azores archipelago. These authors have shown that this seamount is a subsided island, of a typical guyot shape, now at some 440 m below the ocean surface. Carbonates recovered from the top of the guyot are reported to have an age younger than 1.8 Ma and to have been precipitated in water depths that are considerably less deep than their present position indicates. The difference between the present depth and depth of deposition, on average 360 m, is greater than likely fluctuations in sea level, and some seamount subsidence mechanism is required to explain these observations. Ryall *et al.* pointed out that the required subsidence is greater than would be expected from the cooling and contraction of normal oceanic crust alone but they did not offer an explanation for this discrepancy.

Several processes other than thermal contraction may contribute to changes in the depths of seamounts and guyots, for example, the subsidence caused by sediment loading of the surrounding sea floor and of the guyot platform, and the modification of the depth when a seamount rides over the fore-bulge of a subduction zone. Both factors can be significant in certain circumstances (e.g., for the Pratt–Welker seamounts, Lambeck *et al.* 1984) but neither is relevant to the case discussed by Ryall *et al.* (1983). Another possibility is that the anomalous subsidence is a consequence of stress relaxation in the underlying crust and lithosphere. Some authors have argued that the lithosphere responds essentially as an elastic layer when loaded (e.g., Courtney and Beaumont 1983; Watts *et al.* 1980; Turcotte 1979; although the last author was heard to state during the Hamburg International Union of Geodesy and Geophysics (IUGG) meeting that "of course we all know that the lithosphere is mushy"). Others have argued that stress relaxation occurs when the lithosphere is loaded on geological time scales. Thus Walcott (1970), Beaumont (1978), and Lambeck and Nakiboglu (1981) have modelled the lithosphere as a viscoelastic layer and, in the case of the last authors, also as an elastic layer over a viscoelastic layer. England and McKenzie (1982) have modelled the lithosphere as a viscous layer.

Observational evidence for the relaxation time constant is limited. Lambeck (1981*a*) found that the effective flexural rigidity of the oceanic lithosphere in the southeast Pacific decreased with the age of the load and interpreted this in terms of a Maxwell relaxation time of $5-10 \times 10^6$ years. The analysis by Lambeck *et al.* (1984) of the state of isostasy and subsidence of the Pratt–Welker seamounts appears to confirm this interpretation. Much of this and other evidence (e.g., Lambeck and Nakiboglu 1981) is circumstantial because of inadequate geophysical and chronological control. The results by Ryall *et al.* (1983) may therefore provide a useful constraint on this hypothesis.

Subsidence history

Ryall *et al.* (1983) have summarized the known history of the Azores seamount. Their relevant observations are as follows (see also Fig. 1): (i) the crustal age below the seamount is about 10 Ma; (ii) the uppermost basalts were formed subaerially 4.8 ± 0.2 Ma ago; (iii) the age of carbonates near the top of the seamount is ≤1.8 Ma; (iv) these sediments were deposited in a water depth of between 30 and 200 m; (v) the present minimum depth of the guyot is 440 m and the average difference between the depth at time of carbonate precipitation and the present depth has been estimated by Ryall *et al.* as 360 m; (vi) subsidence of the sea floor and the guyot because of thermal contraction of the lithosphere, based on the thermal

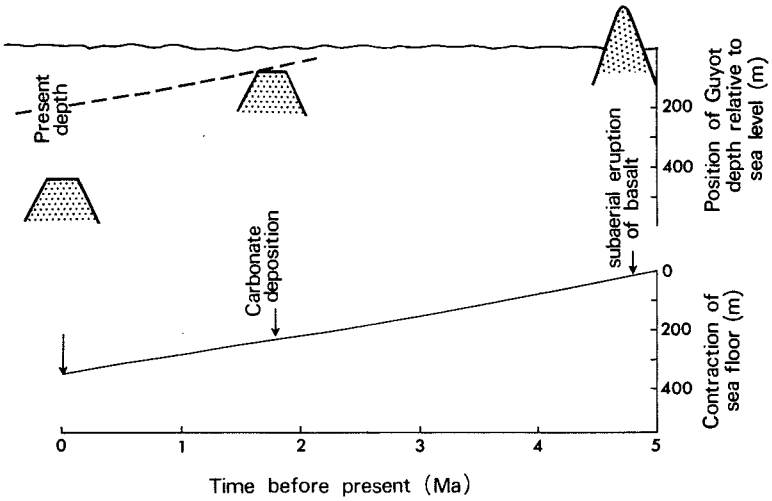

FIG. 1. Schematic illustration of the guyot subsidence with respect to a sea level that may itself vary through time. At 4.8 Ma subaerial basalts erupted; 3 Ma later the guyot was at some depth below the ocean surface and the carbonates were deposited. The guyot is now about 440 m below present sea level. The lower curve represents the subsidence caused by thermal contraction. If this is the sole cause of subsidence, the guyot tops should lie on a parallel curve at times $t = 0$ and 1.8 Ma (e.g., dashed line).

contraction constant of 380 m $Ma^{-1/2}$, is 335 m since the time of basalt eruption 4.8 Ma ago, but only 115 m since the beginning of the carbonate precipitation 1.8 Ma ago; and (vii) sea level has not been constant throughout this time interval. Ryall *et al.* suggested that sea level at 1.8 Ma ago was possibly lower by as much as 100 m.

The lowstand may actually have been lower than this. Vail *et al.* (1977) suggested that in the late Pliocene – early Pleistocene, at the time that the carbonates in question were deposited, sea level was about 150 m below the present level (see their Fig. 6). The discordant subsidence of the Azores guyot is therefore about 100–150 m in a time interval from 1.8 Ma ago to the present. One obvious possibility is that the average thermal contraction constant is too low, but a value of about 700–900 m $Ma^{-1/2}$ is required to explain the discordant subsidence. There appears to be no evidence for this, even for areas close to the ridge axis (Tréhu 1975).

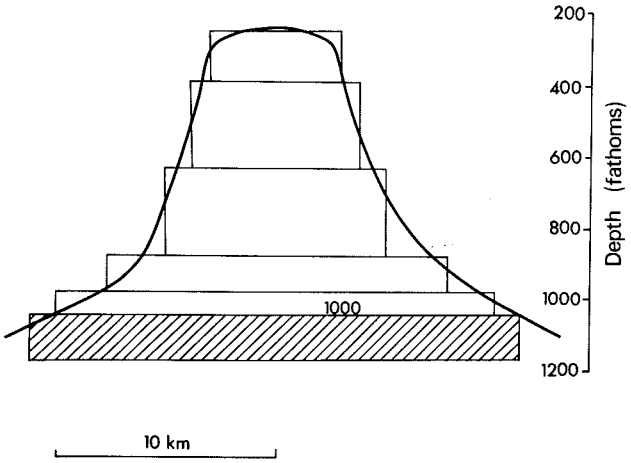

FIG. 2. Load model of the subsiding seamount. Note: 1 fm = 1.8 m.

Stress relaxation

Models for guyot subsidence caused by stress relaxation have been discussed by Lambeck and Nakiboglu (1981) and Lambeck *et al.* (1984). The upper part of the lithosphere is thought to behave mainly as an elastic layer although stress relaxation may occur by macroscopic processes caused by the fracturing of the cooling layer and the presence of water down to considerable depths. The central part of the lithosphere may be predominantly elastic on the time scale in question. The warmer lower part of the lithosphere is believed to be more viscous, and stresses in that layer will relax with time. For the present purpose, the overall response of the lithosphere can be modelled by a viscoelastic layer with an effective flexural rigidity D. An effective viscosity of the subsidence can then be calculated using the formulation of Lambeck *et al.* (1984), once the load is specified. Ryall *et al.* (1983) have surveyed the guyot in some detail. It is approximately axisymmetric, with a radius at its base of about 12 km. We model it by a series of superimposed discs, as shown in Fig. 2. The topography is that seen at present, not the original load at time $t = 0$ required for the relaxation correction. Hence a further disc has to be added of thickness equal to the stress-relaxation subsidence that has occurred in the 4.8 Ma since $t = 0$. This is itself a function of viscosity, and an iterative approach is required. Trial calculations indicate that the total viscoelastic subsidence since the seamount formation is of the order of 200–400 m, and a disc of 300 m has been added. A load density of 2.7 g cm^{-3}, an average crustal density of 2.9 g cm^{-3}, and an upper mantle density of 3.3 g cm^{-3} have been adopted. Any moat created by the initial elastic deformation is assumed to have been filled in by sediments and volcanic debris.

The subsidence curve is a function of the effective flexural rigidity and of the effective viscosity or relaxation time τ. We adopt a Maxwell rheology for which τ = viscosity/rigidity. We also introduce a dimensionless time $t^* = t/\tau$. Young loads on 70 Ma old lithosphere indicate that $D \simeq 3 \times 10^{22}$ N·m (Lambeck 1981*b*). Thermal models of the ocean lithosphere suggest that the effective plate thickness H increases approximately according to $(\text{age})^{1/2}$ if H is defined by a nominal isotherm. Then, using the thermal cooling model of Parsons and Sclater (1977), H for 5 Ma old lithosphere is about one third that of 70 Ma old lithosphere. Furthermore, since $D \propto H^3$, $D_{5\text{Ma}} \simeq D_{70\text{Ma}}/3^3 \simeq 10^{21}$ N·m. Figure 3 gives the subsidence curve

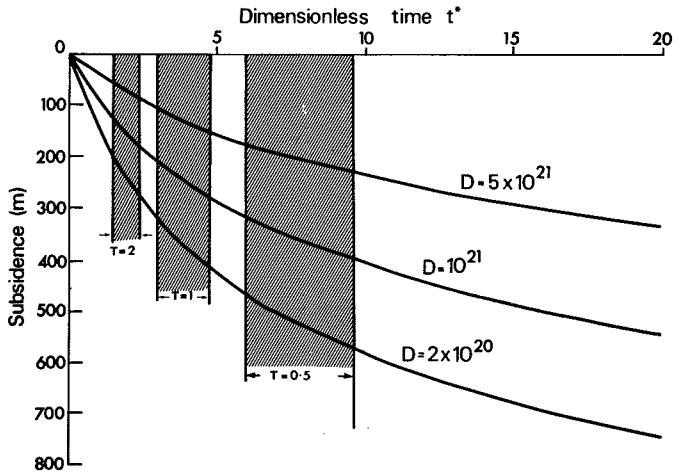

FIG. 3. Subsidence curves caused by stress relaxation for lithospheres of different values for D (in N·m) subjected to the load illustrated in Fig. 2. The relaxation time τ is defined by viscosity/rigidity and the dimensionless time t^* is t/τ. The shaded areas define, for a given value of τ (in Ma) and a given D value, the subsidence that can occur between the time that the carbonates were precipitated and the present. For $\tau = 0.5$ Ma, for example, $t_{\text{initial}} = 3$ Ma and $t^*_{\text{initial}} = 6$, and $t_{\text{final}} = 4.8$ Ma and $t^*_{\text{final}} = 9.6$. The subsidence in the interval is about 110 m for $D = 2 \times 10^{20}$ N·m and about 60 m for $D = 5 \times 10^{21}$ N·m.

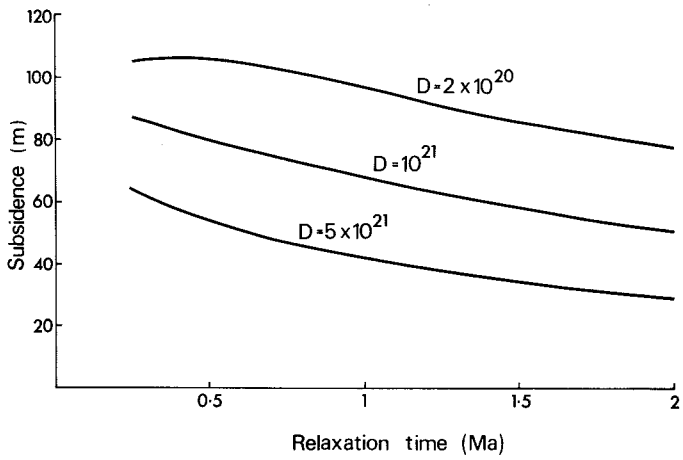

FIG. 4. Maximum subsidence in the interval $t_{\text{final}} - t_{\text{initial}}$ as a function of relaxation time and D (in N·m).

for this and two other values of D. The starting time $t^* = 0$ corresponds to the time of seamount formation. The time at which the carbonates were deposited is then $t = 3$ Ma and the interval of subsidence is from $t = 3$ Ma to $t = 4.8$ Ma. Superimposed on Fig. 3 are some dimensionless time intervals available for the observed subsidence for different τ values. That is $t^*_{\text{initial}} = 3$ Ma/τ and $t^*_{\text{final}} = 4.8$ Ma/τ. The maximum permissible subsidence by this model is about 100 m unless D is much less than 2×10^{20} N·m (Fig. 4). This subsidence value is equal to the lower limit of the observed anomaly.

For a given D value the observed subsidence does not give a unique determination of the relaxation time (see Fig. 4 where this is evident in the result for $D = 2 \times 10^{20}$ N·m). For the load model and plate parameters under consideration here, the lower value is extremely short and improbable. An increase in the load density increases the overall subsidence, as does a higher crustal density or a reduced upper mantle density. Thus the given subsidence values may be about 10–15% higher than given here. The results are quite insensitive to the disc representation of the load but a further 10% uncertainty may come from the fact that a proper iteration for the lowest disc has not been carried out.

For $D \simeq 10^{21}$ N·m as suggested above, the maximum subsidence is about 70–80 m for τ in the range $0.5-1.0 \times 10^6$ years. In view of the above uncertainties this is sufficiently near the lower limit of the anomalous subsidence. If the upper limit of about 150 m is the correct measure of anomalous subsidence it may be necessary to modify some of the ages, so as to increase the time available for stress relaxation. The data of Ryall *et al.* (1983) appear to preclude this. Another possibility is that the total mass of the load has been underestimated but, again, there is no evidence for this in the data presented.

Conclusion

The observations of Ryall *et al.* (1983) indicate that mechanisms other than thermal contraction are required to explain their observed guyot subsidence. A likely mechanism is that of stress relaxation, whereby the flexural stresses created by the load gradually decrease with time. For maximum subsidence the relaxation constant is between about 0.5 and 1.0 Ma for $D \simeq 10^{21}$ N·m. This is for lithosphere aged between 5 and 10 Ma. This compares with about 1 Ma for 20 Ma old lithosphere in the Gulf of Alaska (Lambeck *et al.* 1984) and about 5–10 Ma for 70 Ma old lithosphere in the South Pacific (Lambeck 1981*a*). Lower values for the younger ages are to be expected in view of the thermal evolution of the lithosphere. The present observations do not constrain well the relaxation time. In particular τ is a function of the choice of effective flexural rigidity and other constraints are required. Unfortunately this particular seamount is small and unlikely to give a significant geoid or gravity signal from which D may be estimated independently of the subsidence. If the observation of Ryall *et al.* could be repeated for other guyots in the area then it may be possible to provide a better constraint on the underlying lithosphere's relaxation time.

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