SUBSIDENCE AND FLEXURE ALONG THE PRATT-WELKER SEAMOUNT CHAIN

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ABSTRACT


Chemical and age data led Turner, Jarrard and Forbes to conclude that the origin of the Pratt-Welker seamount chain in the Gulf of Alaska cannot be attributed to a single cause. They argued instead that some seamounts in the chain formed over a hotspot, away from a ridge, while others formed near a ridge. They also noted that the latter group of guyots were generally deeper than the former and they used this observation to predict the origin of the unsampled seamounts in the chain. A second geophysical test of the origin hypothesis is to examine the degree of isostatic compensation of the guyots; seamounts formed near a ridge should be in a state of local isostatic compensation, while seamounts formed away from a ridge should be regionally compensated. This test has been carried out using GEOS 3 and SEASAT altimeter data. The effective flexural rigidity of the lithosphere below all seamounts is found to be less than about $10^{20}$ Nm, such that the isostatic state is nearly local, rather than regional. This may be a consequence of all the seamounts having formed on an initially weak lithosphere, of stress relaxation subsequent to their formation away from the ridge, or both. If the seamounts from Giacomini to Durgin formed away from the ridge then these results point to an effective flexural rigidity at the time of loading of about $3 \times 10^{21}$ Nm and to a stress-relaxation time of about $10^6$ years. These values are for an ocean lithosphere that was about 20–22 my old when loaded. Corresponding values for 60 my old lithosphere in the southern Pacific were previously found to be about $3 \times 10^{22}$ Nm and $5 \times 10^6$ years. This comparison suggests that both the initial elastic response and the rate of stress relaxation are functions of the age of the lithosphere. The subsidence of guyots is due to numerous factors including thermal contraction of the seafloor, sediment loading, the flexure of the lithosphere prior to its subduction along the Aleutian Trench and, in view of the above short stress relaxation time, stress relaxation. A principal uncertainty in evaluating the subsidence that has occurred subsequent to the seamount having been eroded to sealevel is the erosion time interval. The comparison of the predicted subsidence with observed depths points to an erosion time constant of the order of 5 my and which is a function of seamount size. The conclusions from the flexure and subsidence analysis as to where the various seamounts formed are in agreement with those of Turner et al. Giacomini, Quinn, Surveyor, Pratt and Durgin formed away from a ridge and are consistent with a fixed hotspot and uniform spreading rate model. The geophysical information for Denson, Davidson and Hodgkins is consistent with the hypothesis that these guyots formed near or on a ridge. The case for Welker seamount is ambiguous, and this guyot may have formed over a second hotspot, located at an intermediate distance between the first and the ridge. The geophysical evidence for Bowie seamount is also ambiguous. Possibly it has a similar source to Welker, suggesting that there may actually be three different origin mechanisms that led to the chain.

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INTRODUCTION

The Pratt-Welker, or Kodiak-Bowie seamounts, form a submerged chain of seamounts and guyots extending for more than 1000 km from near Queen Charlotte Islands to the Aleutian Trench east of Kodiak Island (Figure 1). Recent geochemical studies and potassium argon dating have led Turner et al. (1980) to suggest that the origin of these seamounts cannot be attributed to a single cause, but that some formed over a hotspot located about 100 km south-east of Bowie seamount while others formed at or near a former spreading ridge. Denson, Davidson and Hodgkins are reported to comprise, in part, basalts that are transitional between ocean ridge and alkali basalts and their ages are comparable with the ages of the crust on which they rest. These seamounts are therefore assumed to have formed near an ocean ridge. Kodiak, Giacomini, Dickins and Bowie are reported to consist of alkali basalts and trachytes. K-Ar ages for this group are 15–25 million years (my) younger than the corresponding crustal age and it has therefore been proposed that these seamounts formed away from a spreading centre. The limited sampling

Fig. 1. Principle bathymetric features of the Gulf of Alaska. Depth contours are in fathoms. The detailed bathymetry is given by Chse et al. (1970) and GEBCO (1979). The dashed line represents motion along a small circle about the Pacific Plate's pole of rotation as given by Turner et al. (1980). Also illustrated are the locations of the two bathymetry profiles given in Figure 11.
of the guyots, however, means that neither the age nor compositional data provides compelling evidence for one origin hypothesis or the other and further tests are desirable. Several geophysical tests of the origin hypotheses are possible. One, used by Turner et al., is to compare the observed guyot depths with those predicted according to either hypothesis of formation. The other is to investigate the degree of isostatic compensation of each seamount, for this will be a function of several factors, including the formation hypothesis.

The first approach assumes that the seamounts are indeed guyots. The Pratt-Welker seamounts are variously called guyots or seamounts on the bathymetric map of Chase et al. (1970). The GEBCO (1979) charts use a similar terminology and give similar depths for the guyot platforms except that for Bowie they give a depth of 440 m as compared with the 235 m given by Chase et al. Herzer (1971) gives a detailed survey for Bowie which indicates terraces at depths of 65 to 100 m with deeper terraces at 220–250 m. It appears from the GEBCO map that only some of the seamounts have been surveyed in detail; particularly Giacomini, Quinn and Welker and, to a lesser degree, Pratt, Surveyor, Bowie and Denson. Surveys of Durgin, Davidson, Dickins and Hodgkins appears to be much less complete. Turner et al. assume that all these volcanoes are guyots and adopt the minimum depth below sealevel as a measure of subsidence. Only some of the seamounts have been dated directly by Turner et al. Of these Kodiak, Giacomini, Dickins and Bowie have ages that correspond to a model of uniform seafloor spreading at a rate of 4.4 cm yr⁻¹ (Turner et al., 1980). Table 1 summarizes these ages.

Turner et al. noted that the depths of the guyots in the first group—Denson, Davidson and Hodgkins—are generally greater than those of the guyots forming the second group—Giacomini, Dickins and Bowie. They use the depth observations of the other guyots in the chain to classify them according to formation near or away from an ocean ridge. In determining to which group a guyot belongs, they compare the observed depth with the depth predicted by the thermal contraction model of an aging lithosphere (Sclater et al., 1971), a process that may continue to be significant for seafloor ages up to about 70 my. From this comparison, Turner et al. conclude that Quinn, Surveyor, Pratt and possibly Durgin, formed away from the ridge. These seamounts, therefore, fall into the second group of Giacomini, Dickins and Bowie (Table 1). The data for Welker seamount do not fit one model better than the other.

Several factors not considered by Turner et al. contribute to the observed variations in depth. Equally important to thermal contraction is the seafloor subsidence caused by the sediment loading because sedimentation rates in the Gulf of Alaska appear to be particulary fast and regionally variable (e.g. Kulm et al., 1973). Another factor is uplift of the seafloor at the outer rise due
Summary of the characterization of the Bowie-Kodiak seamounts according to whether they formed near a ridge (Group 1) or over a hot spot away from a ridge (Group 2) (after Turner et al., 1980). The crust and seamount ages in millions of years are given in parenthesis. Seamount ages given here and in Table 3 are predicted ages based on the assumption of uniform spreading over a hot spot in case of Group 2 and away from a ridge in the case of Group 1. The spreading rates have been estimated from the observed ages of those seamounts that have been dated (indicated by asterisks).

The comparison of the predicted and observed guyot depths gives a measure of subsidence since the time that the seamount was eroded down to sealevel and it is the age of the seamount with respect to this event that is important. The time interval $\Delta t$ between volcano formation and its reduction to sealevel is a function of the subsidence processes already mentioned, of the erosion of the subaerial mass and of any rebound of the seafloor due to the reduce load (see Menard, 1965, Chapter 5, for a discussion of guyot subsidence). Turner et al. suggest that $\Delta t$ may be of the order 3-4 my although they appear to have used $\Delta t = 0$ in calculating the thermal contraction. The time interval may, however, be considerably larger than this and observations of volcanic islands elsewhere (e.g. the smaller islands in the Hawaiian chain) suggest that the erosion time for the larger seamounts in the Pratt-Welker chain may be of the order of 5-6 my. Such a value for $\Delta t$ leads to substantially different estimates for

<table>
<thead>
<tr>
<th>Group 1 Ridge Origin</th>
<th>Group 2 Hotspot Origin</th>
</tr>
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<tbody>
<tr>
<td>Hodgkins* (17, 15)</td>
<td>Bowie (16, 1)</td>
</tr>
<tr>
<td>Davidson* (18, 16)</td>
<td>Dickins* (21, 4)</td>
</tr>
<tr>
<td>Denson* (20, 18)</td>
<td>Giacomini* (43, 19)</td>
</tr>
<tr>
<td></td>
<td>Kodiak* (47, 24)</td>
</tr>
<tr>
<td></td>
<td>Pratt (34, 14*)</td>
</tr>
<tr>
<td></td>
<td>Surveyor (39, 16)</td>
</tr>
<tr>
<td></td>
<td>Quinn (40, 17.5*)</td>
</tr>
</tbody>
</table>

to flexure of the lithosphere prior to its subduction (e.g. Hanks, 1971). This outer rise typically reaches a maximum of several hundred metres at distances of 100-200 km from the trench axis (e.g. Caldwell et al., 1976) and may contribute to the present depths of the platforms of the guyots nearest to the Aleutian Trench. A further factor may be subsidence associated with stress-relaxation in the lithosphere below the seamount load (e.g. Lambeck and Nakiboglu, 1981), a process that could also be significant on time scales of several tens of millions of years. The final factor, considered by Turner et al., is that the sealevel down to which the seamounts are assumed to have been eroded will itself have been variable over geological time with fluctuations reaching more than 100 m (e.g. Vail et al., 1977).
subidence due to thermal contraction, sediment loading and stress relaxation than if \( \Delta t \) is taken to be negligible.

Turner et al’s hypothesis can be tested by evaluating the effective flexural rigidity of the crust below each seamount using gravity or geoid height anomalies. There is a general consensus that volcanic loads, particularly young ones, are supported regionally and that the degree of regional compensation can be characterized by an effective flexural rigidity, \( D_{\text{eff}} \), which will depend, interalia, on the elastic characteristics of the lithosphere. Guyots formed near the ridge on thin and weak lithosphere tend to be locally compensated, guyots formed away from the ridge will be more regionally compensated. The group 1 guyots should therefore have small effective flexural rigidities while the group 2 guyots should have larger values. The duration of loading may complicate any simple relation because the lithosphere under old loads may have been subjected to stress relaxation resulting in reduced \( D_{\text{eff}} \) values (Walcott, 1970; Beaumont, 1978; Lambeck, 1980b).

The objective of this paper is twofold: (i) to test Turner et al’s conjecture that some of the seamounts in question formed over a near-ridge hotspot while others formed away from the ridge, and (ii) to investigate the dependence of \( D_{\text{eff}} \) on the age of the lithosphere and on the duration of loading. In so far as the evaluation of subsidence due to stress relaxation requires a value for \( D_{\text{eff}} \), the geoid observations are analysed first. Guyot subsidence due to the various processes discussed above is then evaluated.

**GEOID ANOMALIES AND FLEXURAL RIGIDITIES**

The geometry of the mean sea surface, taken to be equivalent to the geoid without applying any oceanographic corrections, has been computed for the Gulf of Alaska from GEOS 3 and SEASAT radar altimeter data using procedures that have been previously outlined (Coleman, 1980). The result, Figure 2, has been subjected to some low pass filtering such that anomalies with wavelengths less than about 300 km do not appear. Thus the typical seamount anomalies, of one or two metres in amplitude and 100–300 km in wavelength, are not evident over the Pratt-Welker chain in this figure. Observational accuracies for the GEOS 3 data are variable but, for the passes considered here, it appears to be of the order of ±20 cm. This is based on the comparison of nearly colinear passes and on comparisons of GEOS 3 and SEASAT data for passes over the same topographic features. The distribution of GEOS 3 sub-satellite tracks is dense and a number of satellite passes lie close to the centres of seamounts. The geoid anomalies over these seamounts are compounded by several factors. For south to north passes, the signal near the northern end of the chain is dominated by the trench (Figure 3a), while for north to south passes the signal in the vicinity of the southern seamounts is
dominated by the continental margin (Figure 3b). Other south-north passes lie over or near several seamounts at the same time (Figure 3a). The SEASAT data is more precise but also more sparse in its coverage and only a few passes lie within 10–20 km of the seamounts.

In comparing the geoid anomalies with the bathymetry, it was noted that the maximum anomalies were not always coincident with the coordinates of the seamounts given by Turner et al. This may be due to non-coincidence of the centroid of the seamount with its highest recorded elevation or simply a consequence of errors. We have, therefore, checked the location of each seamount, and relocated its centroid when necessary, using the procedure of Lambeck and Coleman (1982a, b). Giacomini required a correction of about 20 km to Turner et al's coordinates and this brings their location nearer to that given on the map of Chase et al. (1970). For Quinn, Surveyor, Durgin and Denson the correction was less than 5 km.
Fig. 3. Two GEOS 3 altimeter passes in the vicinity of the Bowie-Kodiak seamounts. (a) a southeast to northwest pass near the northern end of the chain and crossing the Aleutian trench. (b) a northeast to southwest pass near the southern seamounts of Hodgkins and Davidson and crossing the continental margin.

The bathymetric maps of Chase et al. have been used to establish best fitting radially symmetric models for each seamount. Most of the seamounts in the chain are well represented by the axisymmetric models, and any errors introduced by this approximation are unlikely to exceed those due to the uncertainty in the bathymetry itself. The seamounts have been continued down from the seafloor to the appropriate depth in the sediment sequence by assuming that their below-sediment slope was identical to their average above-seafloor slope (see below). The downward continuation has also been carried out on different assumptions, e.g. a reduced slope and small variations in the depth of the downward continuation, but these tests indicate that the predicted geoid anomalies are relatively insensitive to the detail of this additional mass. This modification does presuppose an age of the load, and hence an origin, so that the seamount base can be correctly positioned in the sediment sequences. In the first instance, we adopt ages that are based on a continuous age progression from Giacomini to Bowie (Table 1). Figure 4 illustrates the mean bathymetric profiles. The northern group of seamounts from Welker and Giacomini are very comparable in volume and generally larger than the southern group.

For each seamount, a composite geoid anomaly is obtained by plotting
Fig. 4. Mean bathymetric profiles for the seamounts from Giacomini to Bowie. The depths to the base of the seamounts are indicated on the assumption that the seamounts formed over the same hotspot. For Denson, Davidson and Hodgkins, a second depth to the base is shown (dashed lines) and is based on the assumption of a ridge origin.

Geoid height as a function of distance from the centroid. This is permissible here since, for the north-south passes at least, the contamination of one seamount signal by a neighbouring seamount is generally below the data noise level. In constructing these composite profiles, a bias has been subtracted from each individual profile such that at distances beyond 100 km from the centroid, the average height anomaly is zero. Model calculations have indicated that, for a wide range of flexural rigidity values, the anomaly beyond this distance is always less than 10 cm. In some instances the signal is perturbed by known neighbouring bathymetric features and in this case the perturbed part of the signal has been excluded from the bias calculation. For the southern seamounts, the profiles contain the continental margin signal and we have endeavoured to remove this by using a mean margin signal as reference. This has been computed from all north-south passes near the seamount but uncontaminated by the seamount itself. "Topographic noise", ...
contributions from the asymmetry of the load, from topographic features on
the volcano flanks and from uncertainties in the bathymetry, increases with
distance from the load centre (Lambeck, 1981b), from about ±10 cm near the
load centre to as much as 30–40 cm at distances beyond about 50 km when
the topography is particularly asymmetrical.

The results for Giacomini are illustrated in Figure 5. Two composite
profiles have been computed, corresponding to a hypothetical pass over the
centre of the guyot. One profile is based on the 5 GEOS 3 passes (left-hand
side) and the other on the two SEASAT passes (centre). The agreement is
satisfactory even though, as expected, the GEOS 3 profiles exhibit more
variability due to (i) greater noise and (ii) the greater number of GEOS passes
resulting in increased “topographic noise”. The small seamount between
Giacomini and Quinn, for example, may add as much as 30 cm to the signals
of the passes to the south-east of Giacomini. Likewise, Quinn may contribute
some 20 cm to the residuals for the southern-most passes. The composite
profile is characteristic of results for all the other seamounts: a maximum
signal of 1 to 2 metres over the centre of the load, reduced to about zero some
50–80 km away. There is a suggestion of negative flanks of about 20 cm at
about 80–100 km distance in all cases, but it is generally masked by the
topographic noise. Smoothed profiles have been obtained by fitting a fifth order
polynomials through the combined GEOS and SEASAT data points. The fit
has been carried out separately for the two sides of the seamount. Identical
results were obtained using cubic spline functions.

The predicted anomalies are computed using the formalism of Lambeck and
Nakiboglu (1980) for an equivalent elastic plate. The density of the volcanic

Fig. 5. Geoid heigh anomalies over Giacomini Seamount. Left: GEOS 3 composite profile. Centre:
SEASAT composite profile. The solid lines represent the best fit to both sets of data. Right: The mean
observed geoid height and predicted geoid heights on the assumption of local isostasy (curve 1),
$D = 10^{45}, 5 \times 10^{45}$ and (curve 4) $10^{49}$ Nm. The inset illustrates the positions of the subsatellite tracks
relative to the seamont. The bathymetry in the inset is schematic only.
load is taken as 2.5 g cm\(^{-3}\), a value found appropriate for other seamounts (Lambeck, 1981, a,b; Strange et al., 1965; Robertson, 1967, 1970). The adopted sediment density is 1.7 g cm\(^{-3}\). A density contrast of 0.5 g cm\(^{-3}\) is assumed across the base of the ocean 3 layer at a depth of 6 km below the basement. In view of the high sedimentation rates, any depression created by the volcanic load is obviously filled in with sediments. The flexural rigidity \(D\) derived by comparing the model with observations is an equivalent or effective parameter, \(D_{\text{eff}}\), which may be expected to depend on time as well as on other parameters. Viscoelastic models can, in the first instance, be approximated by elastic models in which \(D\) depends on the duration of loading. The essential difference between the two models is that the viscoelastic one predicts small negative geoid anomalies in the distance range of 50–100 km for the present seamounts. Older loads will exhibit more pronounced lows as the moat move inwards and become deeper (see, Lambeck and Nakiboglu, 1981). Because of the various noise sources discussed below, we have used the elastic models only as a basis for comparison.

Comparisons of the observed and predicted profiles to estimate \(D_{\text{eff}}\) present considerable difficulty in view of (i) the relatively low signal to noise ratios, (ii) the contamination of the seamount signal by other geophysical signals and (iii) inadequate coverage of the altimeter passes. Both amplitude and wavelength should be matched and it would be appropriate to compute the observed frequency response and compare this with the predicted spectrum. This has not proved to be successful here because (i) the seamount generated geoid signal is contaminated by signals created by nearly features: the subduction zone, the continental margin, some of the deep channels across the abyssal plane and neighbouring seamounts, and (ii) by the lack of information over the centres of the loads. Instead a subjective approach has been adopted in which the observed curve is visually matched with the predicted curves based on different values of \(D_{\text{eff}}\) (Figure 5). This is acceptable because only very approximate \(D_{\text{eff}}\) values can be estimated from the data. Giacomini appears to be close to a state of local isostatic equilibrium if the depth of compensation is taken to lie at the Moho about 6 km below the basement. In view of the relatively detailed bathymetric survey of this seamount it is unlikely that this low \(D_{\text{eff}}\) results is an artifact resulting from an overestimation of the mass of the seamount. In particular, any incompleteness in the bathymetric surveys are likely to result in an underestimation of the load. Hence the \(D_{\text{eff}}\) estimated from the observed geoid profile will be too high. The results for the other seamounts are illustrated in Figures 6–9 and Table 2. Quinn (Figure 6) and Surveyor (Figure 7) appear to be associated with only a marginally larger \(D_{\text{eff}}\) than is Giacomini, but the regional compensation becomes more pronounced for Pratt (Figure 8) and Durgin (Figure 9). Pratt Seamount is somewhat irregular in shape due to the small topographic
Fig. 6. GEOS 3 geoid height anomalies over Quinn Seamount. The predicted values are for local
isostasy (curve 1), $D = 10^{19}$, $5 \times 10^{19}$ and (curve 4) $10^{20}$ Nm.

Fig. 7. Surveyor Seamount. Same as Figure 6 with predicted curves for local isostasy (curve 1),
$D = 10^{19}$, $5 \times 10^{19}$, and (curve 4) $10^{20}$ Nm.
Fig. 8. Pratt Seamount. Same as Figure 6 with predicted curves for local isostasy (curve 1), $D = 10^{20}$, $5 \times 10^{20}$, and (curve 4) $10^{21}$ Nm.

Fig. 9. Durgin Seamount. Same as Figure 5 with predicted curves for local isostasy (curve 1), $D = 10^{20}$, $5 \times 10^{20}$, $10^{21}$, and (curve 5) $5 \times 10^{21}$ Nm.
Table 2 summarizes the estimates for $D_{\text{eff}}$ for the seamounts from Giacomini to Durgin. The individual estimates are not very precise but of interest is that all values are low and possibly that $D_{\text{Durgin}} \geq D_{\text{Pratt}} \geq D_{\text{Surveyor}} \simeq D_{\text{Quinn}} \geq D_{\text{Giacomini}}$ (1)

Beyond Durgin this trend of increasing $D$ from north-west to south-east does not persist. The composite GEOS 3 profile for Welker indicates a geoid anomaly that is compatible with a state of local isostasy (Figure 10). A similar conclusion has to be reached for the other seamounts, Denson, Dickins, Davidson, Hodgkins and Bowie, although the observational evidence is less reliable because of (i) smaller signal-to-noise ratios and (ii) the complication introduced into the geoid signal by the nearby continental margin (see Figure 3b). The small signal to noise ratio is in keeping with the
small size of these seamounts when compared with the northern-most group (Figure 4). For Hodgkins and Bowie, for example, the maximum observed GEOS 3 geoid anomalies, over the centres of the seamount, do not exceed 75 cm. The few available SEASAT passes that lie within 10 km of these seamounts confirm their nearly local isostatic states.

The results point to a low effective flexural rigidity for the lithosphere underneath the seamounts when these $D_{\text{eff}}$ are compared with results obtained elsewhere. In the South Pacific, for example, $D_{\text{eff}} \approx 3 \times 10^{22}$ Nm for 60 my old lithosphere subjected to short loading times (Lambeck, 1981a). The more typical values of $(1-10)10^{22}$ Nm are clearly excluded by the present geoid data which point to an average $D_{\text{eff}}$ of about $10^{20}$ Nm or less. Only the Durgin value is larger than this but this value is also likely to be overestimated in view of the possible underestimation of the load volume of this guyot. The low values are indicative of a high degree of local isostatic compensation. This may be a consequence of stress relaxation subsequent to the guyot formation, or it may be a consequence of an initially weak lithosphere.

If the seamounts formed over a hotspot away from the ridge, as suggested by Turner et al., then the average age of the lithosphere at the time of loading would be about 20 my (see Table 1). If the lithosphere responded elastically to the load then the appropriate $D_{\text{eff}}$ value would be that for 20 my old lithosphere (c.f. Cazenave et al., 1980; Watts et al., 1980). Thermal models of
the ocean lithosphere suggest that the effective plate thickness $H$ increases about twofold when the age of the lithosphere increases from 20 to 60 my (see, for example, the cooling plate model of Parsons and Sclater (1977) where we define the plate thickness by a nominal isotherm of 500°C). Then, as $D_0 \propto H^3$, $D_0$ for 20 my old lithosphere can be expected to be an order of magnitude less than that for 60 my old ocean, or $D_0 \simeq 3 \times 10^{21}$ Nm. Apart from Durgin, the observed $D_0$ are significantly less than this by a factor of 30 or more (see Table 2). Either the seamounts formed on much younger lithosphere, or there has been stress relaxation following upon formation such that $D_{\text{eff}}$ has decreased with the duration of loading.

An approximate time dependence of $D_{\text{eff}}$ due to stress relaxation with a time constant $\tau$ ($\equiv$ viscosity $\eta$/rigidity $\mu$) is given by Lambeck and Nakiboglu (1981) as

$$\frac{D_{\text{eff}}(t)}{D_0} = \frac{\alpha^4 e^{-\alpha^4 t^*/(1+\alpha^4)}}{(1+\alpha^4) - e^{-\alpha^4 t^*/(1+\alpha^4)}}$$

(2)

where $D_0 = D_{\text{eff}}(t = 0)$, $t$ being the duration of loading. Also $t^* = t/\tau$ and

$$\alpha^4 = \left(\frac{R}{\pi}\right)^4 \frac{\rho_s g}{D_0}$$

$R$ being the radius of an equivalent disc representation of the load, $g$ is gravity and $\rho_s$ is the difference in density between the substratum ($\rho_m$) and water ($\rho_0$). The ratio (2) is largely independent of $D_0$ and, for the seamounts considered here where $R \simeq 40$ km, also of the size of the load. The estimates of $D_{\text{eff}}$ from Giacomini to Durgin are not sufficiently precise for any meaningful estimate of $\tau$ to be estimated from the trend (1), particularly since the Durgin estimate is uncertain. Instead we adopt an average value of $D_{\text{eff}} \simeq 10^{20}$ Nm for this group. This is less than the expected value by a factor of about 30 or more. If this reduction is attributed to stress relaxation then, from equation (2) this relaxation has occurred in a dimensionless time interval of $t^* \simeq 30$. Then, with the duration of loading $t \simeq 12-19$ my, $\tau \simeq (0.5-1.0)$ my. This is a magnitude smaller than that found for the 60 my old lithosphere in the South Pacific (Lambeck, 1981b). An alternative interpretation would be to adopt a higher value for $\tau$ and then predict $D_0$. For example, if a relaxation time of 5 my is adopted, then $D_0 \simeq 5 \times 10^{20}$ Nm and the 20 my old lithosphere would have an effective thickness that is about 25% of that of 60 my old oceanic lithosphere. With just these data it is premature to speculate further except to conclude that both an initially small $D_0$ and a short Maxwell relaxation time $\tau$ appear to be required if these seamounts formed on approximately 20 my old lithosphere.
GUYOT SUBSIDENCE

Thermal contraction

Parsons and Sclater (1977) noted that ocean depths $d_s$ increase with seafloor age $t_s$ according to

$$d_s = a_0 + a_1 t_s^{1/2} \tag{3}$$

with $a_0 = 2500 \text{ m}$, $a_1 = 350$ and the age $t_s$ in units of $10^6$ years. This relation predicts seafloor depths in the North Pacific and North Atlantic for ages less than 70 million years, on the assumption that there has been no sedimentation and concomitant isostatic adjustment of the seafloor. Guyots of age $t_s(\leq t_e)$ will have subsided by amounts

$$d_s(t_e) = d_s(t_s) - d_s(t_s - t_e) = a_1 \left[ t_e^{1/2} - (t_s - t_e)^{1/2} \right]. \tag{4}$$

This is the relation used by Turner et al. (1980) to predict the guyot subsidence in the Pratt-Welker chain. If the depth with respect to sealevel is used as a measure of subsidence then $t_s$ should be replaced by $(t_s - \Delta t)$ where $\Delta t$ is the previously introduced planation constant.

The seafloor bathymetry in the Gulf of Alaska (Chase et al., 1970; GEBCO, 1979) generally increases in depth with age, but not according to equation (3) (Figure 11). This is because sediment thicknesses are of the order of 1 km in many places. The basement topography has been mapped by Silber et al. (1974) and Figure 11 illustrates some comparisons between basement and seafloor topography along two sections parallel to the trend of the seamounts. In a general way the sediments decrease in thickness with distance from the shelf but increase again towards the Aleutian trench. A correction for the deflection of the seafloor by these sediments has been made using a sediment density of $1.7 \text{ g cm}^{-3}$ and a density contrast of $0.5 \text{ g cm}^{-3}$ across the base of the ocean 3 layer where compensation is assumed to occur. The correction is then $0.71H_s$ where $H_s$ is the thickness of sediments. This correction assumes that a state of local isostasy is reached. This is valid when the isopach gradients are small since any creep response of the lithosphere is a function of the wavelength of the load and local isostasy is established relatively quickly for wavelengths greater than several hundred kilometres (Nakiboglu and Lambeck, 1982).

Age estimates of the seafloor along the sections in Figure 11 have been obtained from the magnetic anomaly map of Naugler and Wageman (1973). The variation in the corrected depths with age agree well with those predicted using equation (13), confirming that, once the sediment corrections have been
made, the empirical contraction relation is valid. The predicted depths are generally 100–200 m deeper than the sediment-corrected observed depths and this can be readily attributed to an underestimation of the isostatic factor, a factor of 0.76 being required instead of the above 0.71. Alternatively, this difference can be attributed to renewed expansion of the lithosphere when it passed over the hotspot postulated by Turner et al. (1980) (c.f. Detrick and Crouch, 1978). In either case this difference is inconsequential in subsequent discussions.

With the ages given in Table 1, the subsidence due to thermal contraction subsequent to guyot emplacement follows from equation (4). The results are summarized in Tables 3a and 3b for $\Delta t = 1$ and 5 my. Subsidence is strongly dependent on this choice of $\Delta t$ (compare the results in Tables 3a and 3b).

Fig. 11a. Observed bathymetry profile parallel to and south of the seamount chain, basement bathymetry, sediment corrected bathymetry and predicted bathymetry using a thermal contraction model. The sedimentary sequence is shown in the central part of the figure together with the depth of the seamount base if all seamounts formed over the same hotspot. The corrected less predicted bathymetry, shown in the lower-most part of the figure points to a small rise prior to subduction.
Sediment loading

The total sedimental layer about each guyot can be estimated from the basement map of Silver et al. (1974) and the bathymetry map of Chase et al. (1970). Of interest here is the semimentation that has occurred since the time of guyot formation. Some control on the sedimentation rate is obtained from the Deep Sea Drilling Project (DSDP) site 178. Seismic basement is at about 1000 m depth, whereas coring yielded about 780 m of post-Oligocene (\( \leq 20 \) my) sediments before basalt was penetrated. This probably was not basement rock but a basalt flow coincident with the formation of nearby Giacomini whose age is given by Turner et al. (1980) as a very comparable 20 my. The pre-Miocene sedimentation thickness \( T_{pM} \) at seafloor sites of age \( t_s > t_{Miocene} \) is then taken as...
where $t_{178}$ is the predicted basement age at DSDP site 178 or 50 my, and $T_s$ is the depth to the seismic basement.

Miocene sediments at DSDP site 178 total only about 60 m of the total posy-Oligocene column. At other sites the Miocene sediments are taken as

\[
T_M = \frac{60}{780} (T_s - T_{PM})
\]  
(5b)

if the seafloor age is pre-Miocene, or, for post-Miocene seafloor ages

\[
T_M = \frac{60}{780} (T_s - T_{PM}) \frac{(t_s - t_{Miocene})}{\text{Duration of Miocene}}
\]  
(5c)

Pliocene and Pleistocene sediment thicknesses can be computed in the same manner and it becomes possible to estimate the total sediment thickness at each site at the time of seamount formation. It is the sedimentation subsequent to guyot formation that leads to further guyot subsidence. This can be computed as before (Table 3). The correction is fairly constant for the seamounts between Giacomini and Dickins but decreases rapidly as the seamounts decrease in age. This correction is of the same magnitude and sign as the thermal contraction.

The sediments on the seafloor in the Gulf of Alaska appear to have been largely transported by turbidity currents and significant deposition on the guyot platforms themselves would not be expected. Ice-rafted or glacial erratics are found mainly in the upper hundred metres of sediments in the cores of DSDP sites 177 and 178 (Kulm, 1973) and these may also be expected on the guyot platforms. Their concentration in the upper layers, however, suggests that differential loading from this source is insignificant. Perhaps a more substantial source of differential loading of the guyot platforms is caused by carbonate precipitation above the carbonate compensation level. For typical rates of 10 m in $10^6$ years, Giacomini guyot could have collected up to 200 m of carbonate-rich sediments.

There will be some flexural response to this load. An upper limit to the subsidence may be about 100 m if the response is local. In this case the guyot height will have increased by about 100 m. A lower limit is given by the elastic plate theory and, since the wavelength of this load is quite small, this may be more appropriate. The elastic deflection of the lithosphere to this additional load is given by equation (3a) of Lambeck and Nakiboglu (1980) as approximately

\[
w_0 = \frac{(\rho - \rho_0)}{(\rho_m - \rho_c)} h[1 + a \text{ Ker'} a]
\]
| Seamount   | Observed depth (m)
<table>
<thead>
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<th></th>
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<tr>
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<td>Welker</td>
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<td>Denson</td>
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<td>Hodgkins</td>
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</tr>
<tr>
<td>Bowie</td>
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<th>Seafloor sediment loading (m)</th>
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Notes:
1. hotspot: 1405, ridge: 2345
2. hotspot: 1320, ridge: 1315
3. hotspot: 1320, ridge: 1320
4. hotspot: 1315, ridge: 1315
5. hotspot: 1320, ridge: 1320
6. hotspot: 1280, ridge: 1280
7. hotspot: 1280, ridge: 1280
8. hotspot: 1280, ridge: 1280
9. hotspot: 1280, ridge: 1280
10. hotspot: 1280, ridge: 1280
11. hotspot: 1280, ridge: 1280
## TABLE 3b

Same as Table 3a but for $\Delta t = 5$ my

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<th>Observed depth (m)</th>
<th>Seafloor sediment loading (m)</th>
<th>Guyot sediment loading (m)</th>
<th>Outer rise deflection (m)</th>
<th>Thermal contraction (m)</th>
<th>Stress relaxation (m)</th>
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(7) See note, Table 3a.
(11) Based on guyot ages predicted for the ridge model.
(12) These young seamounts should still be above sealevel if $\Delta t = 5$ my.

### Notes

1. See Figure 1 caption.
2. Chase et al. (1970) give 1207 m. The value of 825 m is given by Turner et al. (1980) and is from a Conrad 1407 track taken subsequently to the Chase et al. charts.
3. This is the depth of the principal platform (see Herzer, 1971).
4. Based on the magnetic anomaly map of Naugler and Wageman (1973).
5. Based on the assumption of uniform spreading and the ages given by Turner et al. for Kodiak, Giacomini and Bowie.
6. Based on the assumption of uniform spreading and the ages given by Turner et al. for Denson, Davidson and Hodgkins.
7. Based on an age given by Menard and McNutt (1982) and attributed to G. B. Dalrymple.
8. Based on guyot ages predicted for the ridge model.
9. With an isostatic factor of 0.71.
10. For an effective guyot age of $t_e = \Delta t$.
11. From Figure 12 with $\eta = 10^{24}$ Pas and $D_0 = 3 \times 10^{21}$ Nm.
where \( a = (\text{radius of load/flexural parameter}) \approx 0.5 \) for this problem. The flexural parameter is defined by equation (8) below. \( \text{Ker}' \) is the derivative of the Kelvin function \( \text{Ker} \), \( \rho \) is the density \( \rho_f \) of the sediment load and \( \rho_c \) is the density of the crust. Then \( w_0 \approx 0.15h \) and the effect of adding 200 m of sediments to the top of Giacomini guyot is to cause a further subsidence of about 30 m. Thus the actual guyot height has increased by 170 m due to carbonate precipitation. All other guyot platforms are assumed to have been modified by an amount that is proportional to their age (Table 3).

**Uplift of the outer rise**

The bathymetry maps do not indicate a pronounced outer rise at the eastern end of the Aleutian Trench but regional geoid maps based on GEOS 3 and SEASAT altimetry data do (Figure 2). The outer rise is clearly defined westward of Kodiak Island (e.g. Watts and Talwani, 1974) where the maximum elevation approaches several hundred metres, in agreement with that predicted by the flexure model of Caldwell et al. (1976). When the bathymetry east of Kodiak is corrected for the differential sediment loading and for the thermal contraction, the residual bathymetry is suggestive of an outer rise with a maximum amplitude of about 200 m at a distance of about 200 km from the trench axis (Figure 11). This corresponds well with the location of the observed outer rise geoid anomaly. Thus Giacomini, Quinn and Surveyor may have been uplifted by as much as 200 m (Table 3).

**Stress-relaxation**

Walcott (1970) recognized that stress relaxation in the lithosphere should be considered when studying the response of this layer to long-term (\( > 10^6 - 10^7 \text{ yr} \)) loads. Watts and Cochran (1974) applied Walcott's formulation to the Hawaiian chain and concluded that there was little observational evidence for stress relaxation subsequent to the volcanic loading. But Beaumont (1978) and Lambeck and Nakiboglu (1981) have pointed out the inappropriateness of this formulation. We have previously pointed to a number of observations that suggest that stress relaxation occurs in the lithosphere (Lambeck and Nakiboglu, 1981). When taken together, the evidence points to a need to investigate nonelastic solutions in more detail than has hitherto been the case, especially when the non-hydrostatic stresses are large and when the time over which the load is applied is long (see also Beaumont, 1979). Yet some, for example Watts et al. (1980) argue that there is no need to invoke stress relaxation when the lithosphere is subjected to stress-differences approaching, according to their model, 1 GPa (10 kbar)
(although Lambeck and Nakiboglu, 1980, have pointed out that such estimates are excessive).

The difference between the two interpretations rests largely on different notions of the rheological zonation of the lithosphere. Advocates of purely elastic models imply that there is a sharp viscosity discontinuity at the base of the elastic layer. We argue that the viscosity gradient, between the mainly elastic and brittle upper part and the more ductile lower part of the lithosphere, is likely to be more gradual. With time, the non-hydrostatic stress migrates upwards into the brittle layer at a rate that will be a function of the dependency of viscosity on temperature, pressure and composition as well as on the detailed mechanisms of creep. In the uppermost part of the cold crust the relaxation of stress may be controlled by mechanical factors such as pore pressure fluid or the behaviour of jointed or fractured rocks. A previous analysis suggested that the time constant for this stress migration is likely to be of the order of $(5-10) \times 10^6$ years (Lambeck, 1981b). In consequence, subsidence of the volcanic loads may be significant over time intervals of $10^7$ years and longer.

Turner et al. (1980) considered that any stress relaxation under the Pratt-Welker seamounts is unimportant on the assumed loading time scale considered here of $(1-2) \times 10^6$ years. They base this conclusion on a time constant for isostatic compensation of the order $10^4-10^5$ years, much less than the formation time of the seamount itself. Detrick and Crough (1978) have used the same erroneous argument. This shorter time scale, is more a measure of the time constant of the asthenosphere or upper mantle, than of the lithosphere. It is this short time constant that permits the introduction of regional isostatic models of an elastic or viscoelastic plate over a fluid substratum if the loading times are much greater than $10^4-10^5$ years. What is of relevance in the island subsidence problem is the possibility of relaxation within the lithosphere.

Lambeck and Nakiboglu (1981) developed the formalism for layered elastic-viscoelastic models but we will use the more simple model here since observationally it is virtually impossible to distinguish between an equivalent viscoelastic layer and these layered models. In their model, the seamount load above the originally undeformed seafloor was kept constant because the deformation under a constant visible load was sought for different epochs. In this study we need to reduce the effective load at any time by the amount of subsidence. A simple approximate solution is available. Consider the equation of vertical equilibrium for an elastic plate over an inviscid fluid, namely

$$D \Delta Aw + \rho_s gw = p$$

where $D = \mu H^3/6(1-v)$ is the effective flexural rigidity of the crust, $\mu$ is the rigidity, $v$ is Poisson's ratio and $H$ is the thickness of the plate. $w$ is the
vertical deformation measured positive downwards, $\rho_s$ is the effective density of the substratum and $g$ is gravity. $p$ is the surface load. If a simple disc load of density $\rho$, radius $A$ and total height $h$, is considered, then for a seamount wholly below sealevel

$$p = g \begin{cases} (h - w_0)(\rho - \rho_0) & r < A \\ 0 & r > A \end{cases}$$

(7)

and

$$\rho_s = (\rho_m - \rho)$$

$w_0$ is the deflection of the plate at the centre of the load, $(h - w_0)$ is the height with reference to the originally undeformed surface and $\rho_m$ is the mantle density. If the moat created by the load is filled in beyond the original depth of the seafloor, equation (2) is still valid provided that

$$\rho_s = (\rho_m - \rho)$$

(see Lambeck and Nakiboglu, 1980). The flexural parameter $l$ is defined by

$$l^4 - \frac{D}{\rho_s g}$$

(8)

The solution of equation (6) with (7) for an incompressible viscoelastic infinite plate loaded by a disc shaped seamount at time $t = t_0$ is (Lambeck and Nakiboglu, 1981)

$$w(r, t) = \frac{\rho}{\rho_s} a \int_{u=0}^{\infty} J_1(au) J_0(ux) W(t, h_a) \, du$$

(9)

where $a = A/l$, $x = r/l$, and

$$W = h_a \left[ 1 - \frac{u^4}{a + u^4} \exp\left[ -(t - t_0)/\tau(1 + u) \right] \right] H(t - t_0).$$

(10)

$J_0$ and $J_1$ are Bessel functions. $H(t - t_0)$ is the Heaviside step function. The apparent height $h_a = (h - w_0)$ remains constant with time (that is, the seamount actually increases in height at a rate equal to subsidence). An appropriate solution to the more realistic problem where $h_a$ decreases due to subsidence can be obtained by writing

$$h_a(t_k) - h_a(t - t_0) = \sum_{n=0}^{k-1} w_0(t_n) \left[ H(t - t_n) - H(t - t_{n+1}) \right]$$

where

$$t_k = t_0 + k\delta$$

$\delta$ being a small time increment, and $w_0(t_k)$ being the deflection at the centre at
time $t_k$. The deflection is obtained by superimposing equation (10) for different Heaviside functions. That is, the solution is given by (9) with

$$W(t_{k-1} < t < t_k) = [h_t - h_0(t_{k-1})] \frac{1}{1 + u^4} + h_t \frac{u}{1 + u^4} (1 - e^{-k\delta})$$

$$- \frac{u^4}{1 + u^4} \sum_{k=0}^{k-1} w_0(t_n) \exp \left[-\frac{(k - 1 - n)\delta - t_0}{(1 + u^4)}\right]$$

$$- \exp \left[-\frac{(k - n)\delta - t_0}{(1 + u^4)}\right] \right\}$$

(11)

If $\delta$ is small compared with $\tau$ then this approximate solution gives satisfactory results. Figure 12 illustrates the subsidence predicted for a characteristic guyot

Fig. 12. Viscoelastic deformation for a lithosphere of given $D_0$ as a function of the dimensionless time $t^*$. Curve a is for $D = 6 \times 10^{22}$ Nm and the initial elastic deflection $w_0$ is 0.31 km. Curve b is for $D = 10^{22}$ Nm ($w_0 = 0.81$ km) and curve c is for $D = 3 \times 10^{21}$ Nm ($w_0 = 1.38$ km).
in the Pratt-Welker chain for different values of \( D_0 \). Note that \( D_0 \) is actually the \( D_{\text{eff}} \) at a time corresponding to the end of loading.

The size of the seamount above the undisturbed seafloor at time \( t = 0 \) (immediately after the initial elastic deflection) is the same in all cases, although the physical load increases with decreasing \( D \) because the initial elastic deflection becomes greater. Hence the solutions for different \( D \) converge on different isostatic states. The strongest plate (curve a) has an initial elastic deflection that is small and the subsequent relaxation is slow. For the weakest plate (curve c) the initial deflection is already close to the local isostatic state and any subsequent relaxation is rapid. For \( D_0 = 3 \times 10^{21} \text{Nm} \) and a viscosity of \( 10^{24} \text{Pas} \) (\( \tau \sim 10^6 \text{years} \)), as suggested by the above geoid analysis, the total subsidence of Giacomini is about 700 m. If \( \Delta t = 10^6 \text{years} \), the subsidence since the time the seamount was reduced to sealevel is about 500 m but if \( \Delta t = 5 \times 10^6 \text{years} \), the post-planation subsidence is reduced to about 160 m. Like the thermal contraction, the subsidence induced by stress relaxation is strongly dependent on this planation time constant. If \( \eta \sim 10^{25} \text{Pas} \) (\( \tau \sim 10^7 \text{years} \)) and \( D_0 \approx 3 \times 10^{21} \text{Nm} \), the total subsidence for Giacomini is about 380 m but the subsidence since planation is still about 160 m for \( \Delta t \sim 5 \times 10^6 \text{years} \). Table 3 gives the predicted subsidence due to stress relaxation since the time that the seamount was reduced to sealevel. The effective viscosity for the lithosphere is taken as \( 10^{24} \text{Pas} \) and the planation time constant is taken as 1 my (Table 3a) or as 5 my (Table 3b).

Changes in the rheological properties due to the thermal evolution of the lithosphere with time have been ignored in these calculations and viscosity values are time and space averaged quantities. Such variation can be resolved in principle by estimating the effective viscosity from differences in subsidence for successive pairs of guyots. But this will only be successful if there are no other factors contributing to the present depths of the guyots and if the planation time constants are adequately known. This is certainly not the case for the Pratt-Welker seamounts.

Discussion

The observed subsidence summarized in Table 3 is based on the minimum platform depth observed over the seamount. This has the disadvantage that the shallowest depth of a guyot may not be recorded on a ship track. An alternative is to use shelf breaks as an indicator of depth. This has been argued by Menard and McNutt (1982) but such a choice has several disadvantages. (i) Conventional bathymetric surveys, digitized at 6 minute intervals, give poor resolution; (ii) multiple terraces of different ages are frequently present; (iii) the timing of terrace formation is generally unknown; and (iv) the terraces
may have formed in times of rapid subsidence during the seamount formation. The subsidence calculations are summarized in Table 3 for $\Delta t = 1$ my. For many of the guyots the predicted total subsidence greatly exceeds the observed depths if it is assumed that all seamounts formed over the same hotspot. More significant than total subsidence is the discrepancy between the observed and predicted change in depth from one guyot to the next, a quantity that is largely independent of sealevel fluctuations, of the constant $a_0$ in the thermal contraction model (equation 3) and of the isostatic factor in the sediment loading calculation. For those seamounts that are postulated to be of ridge origin, the predicted depths are also excessive. Either the principal contributions to the total subsidence are in error, or some of the underlying assumptions are invalid. The essential correctness of the sediment loading effect is confirmed by the fact that the corrected seafloor bathymetry follows the predicted depths based on the global models of thermal contraction of oceanic lithosphere. Also, the depth of the sediment layer since guyot formation is controlled for Giacomini by the seismic result and the DSDP cores. It is unlikely that the estimates for the other seamounts are in error by an order of magnitude. The corrections for the loading of the guyots themselves are relatively inconsequential but also questionable. Diatom sedimentation on these high-latitude guyots may be high. At DSDP site 192, on top of Meiji Guyot, rates of up to 50 m/my have been recorded (Creager et al., 1973). Currents, on the other hand, may remove much of these sediments.

The thermal contraction estimates are unlikely to be in significant error unless the ages of the seamounts are seriously wrong. The uplift of the seafloor near the trench may be in error by 100% but this will not affect the general result either. Subsidence caused by stress relaxation cannot be inconsequential if the effective viscosity of the lithosphere is of the order $10^{24}$ Pas as suggested by the analysis of the geoid data. But even if a much higher viscosity is assumed, the agreement between observed and predicted subsidence is not improved. The source of greatest uncertainty is the choice of $\Delta t$.

For a planation time constant of 5 my the agreement between observed and predicted depths, based on the hotspot hypothesis, is generally better (Table 3b) for those seamounts older than $\Delta t$. The predicted depths, based on the hotspot hypothesis, of the guyots from Giacomini to Durgin now lie within about 200 m of the observed depths. In view of the uncertainties in the various estimates and adding the sealevel fluctuations, this agreement is satisfactory. Of these seamounts, Pratt is the smallest and the predicted depth may be too shallow. This could be indicative of $\Delta t$ being less for Pratt than for the larger Giacomini, Quinn and Surveyor guyots. Clearly, the uncertainty in this constant and its variability from one guyot to the next, makes the predicted depth an unreliable quantity.

Welker seamount has a predicted depth that is less than the observed depth
unless $\Delta r < 5 \text{ my}$. Welker is intermediate in size between Durgin and Giacomin and the depth can be reconciled with a hotspot origin if $\Delta t \sim 2\text{-}3 \text{ my}$. Menard and McNutt (1982) attribute an age for Welker of 15 my to G. B. Dalrymple. This is about 5–6 my older than if this seamount was part of the same hotspot sequence as Durgin to Giacomini but this age is compatible with the observed subsidence if $\Delta r \sim 5 \text{ my}$ (see Table 3b). The predicted depths for Denson are too shallow to be compatible with the hotspot model unless $\Delta r \sim 0$. A near ridge origin may be more appropriate. The depth of Dickins is commensurate with the hotspot origin if $\Delta t \sim 2 \text{ my}$. Davidson is commensurate with the ridge model, as is Hodkins, provided that $\Delta t$ approaches 5 my. This may be incompatible with the previous suggestion that small seamounts should have small $\Delta t$'s. If Bowie formed some 14 my ago near a ridge, its present depth should be comparable to that of Hodkins or Davidson (Table 3b). Instead, the shelf breaks near 80 and 235 m indicate that subsidence has not been excessive and that the guyot must therefore be relatively young.

A comparison of our conclusions with those of Turner et al. (1980) reveals little difference, despite the complications introduced by the additional mechanisms that cause subsidence or uplift of the guyots. That there more complete subsidence model can still be reconciled with the observed depths is a consequence of the introduction of a planation constant that is largely unknown but which may be dependent on the seamount size.

ORIGIN OF THE PRATT-WELKER CHAIN

Turner et al.'s conclusions on the seamount origins are based on the differences between crust and seamount ages and on the magma types of dredged samples. Alkali basalts and trachytes were assumed by Turner et al. to be characteristic of mid-plate volcanisms, while the transitional basalts were assumed to be more characteristic of ridge volcanism. The age results for Kodiak, Giacomini, and Dickins appears to be unambiguous, with whole rock and plagioclase K-Ar ages that are concordant with fission track ages (Turner et al., 1973, 1980). The age for Giacomini agrees with the ages of nearby lava flows encountered at DSDP sites 178 and 179. Kodiak and Giacomini are younger than the crust upon which they rest by about 21–24 my. The differences between the ages for these two seamounts is in general agreement with what would be predicted from a fixed hotspot model and estimates of the Pacific plate velocity. The Dickins result based on samples dredged from two different depths, also point to an age that is distinctly younger than that of the seafloor, by about 16 my.

Both Denson and Davidson have ages that are comparable with the seafloor age. For the former guyot, two samples yield distinctly different ages. 19.7 my
and 8.1 my. The latter age is interpreted by Turner et al. as a minimum age. Hodgkins presents a particular problem, with three different ages being obtained \( \approx 2.7 \) my, \( \approx 6.6 \) my (considered a minimum age) and 14.0 my for the deeper dredge haul. The oldest sample is a transitional basalt, normally characteristic of ridge volcanism, while the other samples correspond to alkali basalts. Turner et al. conclude that these younger ages may reflect a period of late volcanism. Evidence for distinct phases of volcanism, separated by as much as 10 my, if often seen elsewhere; for example, in the Cook-Austral region (compare the results of Dalrymple et al., 1975 with those of Duncan and McDougall, 1976). One is tempted to ask whether an additional dredge haul on Dickins would have led to a similar conclusion. One is certainly led to question the assumption that all volcanism took place is a relatively short time interval.

Turner et al.'s estimate of the age of Bowie seamount, as greater than \( 7 \times 10^5 \) years, is based on the observation that much of the seamount is reversely magnetized. This is not a very useful estimate. They quote a much younger date for a pinnacle rising above the guyot (see also Herzer, 1971) but, in view of the general guyot morphology, this may represent a late phase of volcanism which is not indicative of the age of the seamount as a whole. Turner et al. indicate that the magma types, presumable referring to the pinnacle, are indicative of mid-plate volcanism but again, it is not clear whether this is representative of the guyot as a whole. According to Herzer (1971), intermediate type basalts were also obtained.

Turner et al.'s conclusions about the guyot depths are largely substantiated by our analysis, although we would be less equivocal than they are about adding Durgin and Welker to the group of hotspot-origin guyots. The age reported by Dalrymple of Welker seamount is older than predicted by the hotspot model but, in view of the uncertainty in \( \Delta t \), we cannot use the guyot depth to distinguish between the two ages.

The geoid height analyses for the effective flexural rigidity may also permit a discrimination to be made between the two origin-hypothesis although in the case of the Pratt-Welker seamounts this has not been overly successful. Taken alone, the low \( D_{\text{eff}} \) for the northern group of seamounts from Giacomini to Durgin points to either significant stress relaxation or to formation over a ridge. Welker has a much reduced \( D_{\text{eff}} \), when compared with that of Pratt of Durgin, pointing to formation on a relatively young crust. This suggestion is borne out by Dalrymple's estimated age; for the more northern seamounts \( t_s - t_g \approx 20-24 \) my but for Welker this difference is 11 my. The effective flexural rigidities for the seamounts south of Welker are low but also poorly determined as the seamounts themselves are small. These low values, of the order \( 10^{20} \) Nm, suggest that a state of local isostasy has been attained and, in the case of Denson, Davidson and Hodgkins, this is compatible with the
hypothesis that they formed near a ridge. The result for Bowie is unexpected for it this guyot is as young as assumed by Turner et al., it should be regionally compensated. But the age of the seafloor near Bowie is about 15 my and defining, as before, a nominal effective lithospheric thickness by the depth of the 500°C isotherm in the Parsons and Sclater (1977) model, we find that $H(t=20\text{ my})/H(t=15\text{ my}) \simeq 1.4$ and $D_0(t \simeq 20\text{ my})/D_0(t \simeq 15\text{ my}) \simeq 2.7$. With $D_0(t \simeq 20\text{ my}) \simeq 3 \times 10^{21} \text{ Nm}$, the effective flexural rigidity of the crust below Bowie should be about $10^{21} \text{ Nm}$. Stress relaxation will initially be rapid for such a weak lithosphere (see Figure 12), and the effective flexural rigidity for a 1 million year old seamount may not be so different from that observed. The argument for Bowie could also be used to conclude that Davidson or Hodgkins were young seamounts and the geoid height anomalies do not provide sound constraints on hypothesis of origin for this southern group of seamounts.

The revised guyot depth and geoid height analyses confirm in a general way the hypothesis by Turner et al. That the Pratt-Welker seamount formed in two different locations. The northern group, from Kodiak to Durgin, appear to have formed away from the ridge and the various pieces of evidence for this are consistent. A reasonable argument can also be drawn up for a ridge origin of Denson, Davidson and Hodgkins. Welker appears anomalous in that it does not fall into either category. Bowie also appears to be anomalous in that its position suggests that it should form part of the Denson, Davidson and Hodgkins sequence but the much shallower platform depth of Bowie relative to Devidson and Hodgkins points to a different origin. Possibly Bowie may have been subjected to a later phase of volcanism as it passed over the hotspot, possibly the one that created the northern part of the chain or alternatively, over the one that led to Welker's formation. Only extensive and detailed chemical analyses and K-Ar dating can throw further light on this possibility.

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