A MECHANICAL MODEL FOR PLATE DEFORMATION ASSOCIATED WITH ASEISMIC RIDGE SUBDUCTION IN THE NEW HEBRIDES ARC

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ABSTRACT


Tectonic features associated with a subducting fracture zone-aseismic ridge system in the New Hebrides island arc are investigated. Several notable features including a discontinuity of the trench, peculiar locations of two major islands (Santo and Malekula), regional uplift, and the formation of a basin are interpreted as a result of the subduction of a buoyant ridge system. The islands of Santo and Malekula are probably formed from an uplifted mid-slope basement high while the interarc basin of this particular arc is probably a subsiding basin instead of a basin formed by backarc opening. The situation can be modeled by using a thin elastic half plate overlying a quarter fluid space with a vertical upward loading applied at the plate edge. This model is consistent with topographic and geophysical data. This study suggests that subduction of aseismic ridges can have significant effects on tectonic features at consuming plate boundaries.

INTRODUCTION

Previous workers (Vogt, 1973; Vogt et al., 1976; Kelleher and McCann, 1976; Chung and Kanamori, 1977, 1978) have suggested that the process of plate subduction and the geometry of plate boundaries can be affected by subduction of aseismic ridges because of the associated buoyant force. Vogt (1973) and Vogt et al. (1976) found that aseismic ridges on oceanic plates often trend into cusps or irregular indentations in the trace of the subduction zone, and a region of reduced seismicity seems to be associated with many consumed ridges. Kelleher and McCann (1976) studied the distribution of great earthquakes in the Circum-Pacific belt and found that where aseismic ridges or bathymetric highs interact with active trenches, large shallow earthquakes occur less frequently and have generally smaller rupture lengths than large events along adjacent segments of the plate boundary. Chung and Kanamori (1977, 1978) found that the subduction process, earthquake
mechanisms, and seismicity patterns are significantly different in a region where an aseismic ridge-fracture zone meets the New Hebrides Trench. At the extension of the aseismic ridges of the D’Entrecasteaux (ridge-fracture) zone, intermediate depth earthquakes have shallower focal depths than those occurring on either side. There is also some suggestion that the dip angle of the Benioff zone of the section where the aseismic ridge is subducting is smaller than in the adjacent sections. These can be explained in terms of local uplift of the Benioff zone due to the buoyant force associated with a low density ridge. In this paper we investigate the nature of plate interaction and deformation at the boundary of aseismic ridge subduction in the New Hebrides island arc by using a simple thin elastic plate model. We first summarize various tectonic features of the New Hebrides Arc.

REGIONAL SETTING AND TECTONIC ANOMALIES OF THE NEW HEBRIDES ISLAND ARC

The most remarkable feature of this island arc is perhaps the discontinuity of the New Hebrides Trench between 14.5°S and 16.5°S (Figs. 1 and 2) where a transverse feature called D’Entrecasteaux fracture zone (Mallick, 1973; Luyendyk et al., 1974), with a topographic ridge on each side of it, abuts the island arc. Earthquake hypocenters extend to depths of at least 200 km below the entire length of the arc and there is no diminution of seismicity beneath the trench gap, suggesting that the two trench sections belong to a single subduction zone or a single arc system. Extensive uplift is observed at the section including the Santo, Malekula, Maewo and Pentecost islands. Some areas have undergone Quaternary uplift of at least a few hundred meters (Mitchell, 1968; Coleman, 1970; Mitchell and Warden, 1971). During the earthquake swarm of August, 1965, the largest shocks of which were of magnitude 7 or larger, an uplift of 50–80 cm in the northern part of the island of Malekula occurred along about 100 km of the coastline (Benoit and Dubois, 1971). Near the trench gap there is an abrupt change in the spatial distribution of islands and in the structure of the arc. The two largest islands, Santo and Malekula, deviate from the general trend of the arc and are located at a position which should be a part of the trench if the trench were continuous (Figs. 1 and 2). To the north and south of the trench gap the island arc appears to be a single arc but within the area it becomes a multiple arc. East of Santo and Malekula, there is a basin which is well developed compared with other depressions of this arc; no comparable basins occur elsewhere in the New Hebrides island arc.

The geology of the New Hebrides islands is relatively well known. Details about the geology as well as tectonics of the island arc have been studied and summarized by Mitchell (1970), Mitchell and Warden (1971), Karig and Mannerick (1972), Mallick (1973), Luyendyk et al. (1974), Colley and Warden (1974), and Dugas et al. (1977). The islands of the New Hebrides can be divided roughly into three belts: a western belt consisting of the
Torres Islands, Santo and Malekula; an eastern belt comprising the Maewo and Pentecost islands; and a central chain including the Banks Islands, Aoba, Ambrym, Epi, Efate, Erromango, Tanna, and Aneityum. The central chain includes all the active and most of the recently extinct volcanoes. The islands of the New Hebrides consist primarily of volcanic, volcanoclastic and recent

![Map of the New Hebrides and related features.](image)

*Fig. 1. The New Hebrides island arc and some related features. Submarine contours are in kilometers. ▲ represents historically active volcano.*
Fig. 2. Bathymetry of the New Hebrides island arc and its vicinity. Contour interval is 0.5 km (after Chase, 1971).

sedimentary rocks. The volcanism in the western belt is mainly pre-middle Miocene; the eastern belt is mainly late Miocene to early Pliocene; the central belt is mainly Pliocene to Recent. The periods of volcanism among the belts overlap slightly.

The concept of an extensional origin of inter-arc basin or back arc opening was developed by Karig (1970, 1971a, b) based on studies of the Tonga-Kermadec and the Mariana arc systems. Based on the similarity in tectonic positions, Karig and Mannerickx (1972) proposed that the basin in the New Hebrides is of extensional origin. According to Luyendyk et al. (1974), however, no well-developed interarc basins comparable to the Lau Basin of the Tonga arc system are present in the New Hebrides. Marine geophysical data indicate that more than 1 km of undisturbed sediment is present; no axial high is found and the central basin lacks steep walls. Regional dips of sediment layers in the basin are toward the basin center rather than subhorizontal or outward-dipping as would be expected if rifting is occurring (or occurred). Thus an alternate mechanism for the origin of the interarc basin seems necessary. In this paper, we suggest that this well developed basin is related to subduction of the D'Entrecasteaux ridge rather than by back-arc extension.

A MECHANICAL MODEL OF ASEISMIC RIDGE SUBDUCTION

Chung and Kanamori (1977, 1978) suggested, on the basis of various seismological data, that the buoyant D'Entrecasteaux ridge-fracture zone is underthrusting beneath the New Hebrides Arc. The plate interaction near the subduction boundary is schematically shown in Fig. 3. When a ridge is subducted the upward loading on the upper plate is larger than usual because of the buoyant force. The ridge resists subduction and interacts strongly with
the overriding plate. As a consequence, the large upward loading lifts up the trench floor and the edge of the overriding plate. The islands of Santo and Malekula were probably formed by an uplifted portion of the outer arc or the trench slope break (Dickinson, 1973; Karig, 1974), which is located between the trench and the volcanic arc. This explains the interruption of the New Hebrides Trench, the observed uplift in the area, and the location of the two large islands which are off the general trend of the New Hebrides Arc. As illustrated by Fig. 3, the upward force applied on the overriding plate by the ridge also causes subsidence some distance from the boundary of the overriding plate. Similar subsidence associated with uplift is observed in other situations. For instance during thrust earthquakes along subducting plate boundaries, such as the great Chilean earthquake of 1960, the Alaskan earthquake of 1964 (Plafker, 1972), and the Nankaido earthquake of 1946 (Fitch and Scholz, 1971), uplifts were observed on the seaward side while subsidence occurred in a broad zone on the landward side. A similar situation is the subsidence and uplift associated with glacial loading, at the site of the downward loading subsidence occurs but some distance away uplift occurs (Haskell, 1935; Walcott, 1970; Sweeney, 1977). Another example is the deflection of the oceanic plate near the trench. At the trench downward deflection occurs while at a distance of the order of 100 km seaward an outer rise can be observed. (Hanks, 1971; Watts and Taiwani, 1974; and Caldwell et al., 1976). This kind of trench-outer rise profile has been interpreted as the deflection of a semi-infinite plate, overlying a fluid quarter space, on which vertical and horizontal loads are applied at the edge. Parameters such as effective (elastic) plate thickness, flexural rigidity and the magnitudes of vertical and horizontal loading can be estimated by modeling these processes. In the following we model the overriding lithosphere by a semi-infinite elastic thin plate and the underlying asthenosphere (and the lower lithosphere) by a fluid quarter space.

The geometry of the present problem is illustrated schematically in Fig. 4. The edge of the semi-infinite plate is locally subject to vertical upward load-
Fig. 4. Deflection of a thin elastic half plate over a fluid quarter space under horizontal and vertical loadings \(N_b\) and \(P_b\) at the plate edge. \(x_{m1}\) is the location of the first relative minimum on the deflected surface.

…ing due to the buoyant force associated with the low density ridge. This model is adequate if the time of loading is much longer than the relaxation time of the asthenosphere and much shorter than that of the lithosphere. We simplify the three dimensional problem to a two dimensional problem, and the upward loading to an upward line force at the plate edge \(x = 0\), Fig. 4. The notation employed in this paper is similar to that used by Hanks (1971) and is given in Notation I. For the bending of a thin elastic half plate over a fluid quarter space, in a constant gravity field, the deflection \(y\) satisfies (Hetenyi, 1946):

\[
D \frac{d^4y}{dx^4} - N_b \frac{d^2y}{dx^2} + K_0 y = 0
\]

(1)

Under the boundary conditions:

\(x = 0\): \(Q_v = P_b\); horizontal loading = \(-N_b\); \(M = 0\)

\(x \to \infty\): \(y \to 0\), \(M \to 0\)

the solution to (1) is given by:

\[
y = \frac{P_b}{\beta K_0} \frac{2\lambda^2}{3\alpha^2 - \beta^2} e^{-\alpha x} \left[ 2\alpha \beta \cos \beta x + (\alpha^2 - \beta^2) \sin \beta x \right]
\]

We investigate a case of zero horizontal loading at the plate edge first and then we shall discuss the effect of horizontal loading later in this section. For \(N_b = 0\), we have:

\[
y(x) = \frac{2\lambda P_b}{K_0} e^{-\lambda x} \cos \lambda x
\]

(2)

\[
y(0) = \frac{2\lambda P_b}{K_0}
\]

(3)
NOTATION I

Explanation of the symbols used in this section

\( P_b \) = vertical force/unit width, \( \text{dn cm}^{-1} \)

\( N_b \) = horizontal force/unit width, \( \text{dn cm}^{-1} \) (negative if compressional, positive if tensi-

\( \gamma \) = plate deflection, cm

\( M \) = bending moment/unit width, \( \text{dn} \); \( M = -D \frac{d^2\gamma}{dx^2} \)

\( Q_v \) = vertical shearing force/unit width, \( \text{dn cm}^{-1} \); \( Q_v = \frac{dM}{dx} + N_b \frac{dy}{dx} \)

\( \rho_m \) = density of mantle underlying the elastic plate = 3.4 g cm\(^{-3} \)

\( \rho_a \) = density of air overlying the elastic plate

\( K_0 \) = foundation modulus = \( (\rho_m - \rho_a)g = \rho_m g \), \( \text{dn cm}^{-3} \)

\( g \) = gravitational acceleration = 980 cm s\(^{-2} \)

\( D \) = flexural rigidity, \( \text{dn cm} \)

\( D = \frac{Eh^3}{12(1 - \nu^2)} \)

\( E \) = Young's modulus = \( 6.5 \cdot 10^{11} \) \( \text{dn cm}^{-2} \)

\( h \) = effective (elastic) thickness of plate = \( 25 \cdot 10^5 \) cm

\( \nu \) = Poisson's ratio = 0.25

\( \lambda \) = \( \sqrt[4]{\frac{K_0}{4D}} \), cm\(^{-1} \)

\( \alpha \) = \( \sqrt[4]{\lambda^2 + \frac{N_b}{4D}} \), cm\(^{-1} \)

\( \beta \) = \( \sqrt[4]{\lambda^2 - \frac{N_b}{4D}} \), cm\(^{-1} \)

\( x_{mn} \) = distance from the plate edge to the \( n^{th} \) relative extremum of deflection

The \( n \)th extremum of plate deflection occurs at:

\( x_{mn} = (n - \frac{1}{4}) \frac{\pi}{\lambda}, \quad n = 1, 2, 3, ... \)

For the first one we have (Fig. 4):

\[ x_{m_1} = \frac{3\pi}{4\lambda} \tag{4} \]

and

\[ y(x_{m_1}) = \frac{2\lambda P_b}{K_0} \exp \left( -\frac{3\pi}{4} \right) \cos \left( \frac{3\pi}{4} \right) \tag{5} \]

We now calculate \( x_{m_1}, y(0), \) and \( y(x_{m_1}) \) by employing some reasonable

values for the parameters involved. Taking \( E = 6.5 \cdot 10^{11} \) \( \text{dn cm}^{-2} \), \( \nu = 0.25, \)

\( h = 25 \) km (Caldwell et al., 1976), we have \( D = 9.03 \cdot 10^{29} \) \( \text{dn cm}, \lambda = 1.743 \cdot \)
$10^{-7}$ cm$^{-1}$, and $x_{m1} \approx 135$ km (Fig. 4). Figure 5 shows two topographic profiles, one across line $AA$ and one across line $BB$ of Fig. 2. The overall topography is quite complicated with the northern part of the basin wider than the southern part (profiles $AA'$ and $BB'$ in Fig. 5). We assume that the basin center is the first relative minimum corresponding to that in Fig. 4 and treat the horizontal distance from the highest point in the Santo or Malekula island to the basin center as $x_{m1}$ (Fig. 5). For profile $AA'$ we have $x_{m1} = 118$ km which is consistent with the calculated value 135 km given above. For profile $BB'$, we obtain $x_{m1} = 73$ km which is about half the calculated value. We next calculate $P_b$, the total buoyant force, and $y(0)$, the maximum uplift. Since $P_b = V \cdot \Delta \rho \cdot g$, we need to know the volume of the ridge system, $V$, and the density contrast between ridge and oceanic plate in order to calculate $P_b$. Figure 6 shows the topographic profile across line $CC'$, of Fig. 2, which is roughly perpendicular to the trend of the ridges. The horizontal scale, or the width, of the ridge system across this profile is about 80 km, but the thickness of the ridge is unknown. We assume that the thickness is one half the width of the ridge. The buoyant force is partially from the portion of ridge under the island arc and partially from the portion near the plate boundary but not beneath the island. If the ridge has gone down to 100 km in depth with a dip angle of 60$^\circ$, the down-dip length of the subducted ridge is 120 km. We further assume that part of the buoyant force comes from a 40 km long unsubducted ridge. Then we can calculate the volume of the ridge. The density contrast between ridge and oceanic plate can be between 0.05 and 0.08 g/cm$^3$ (Kelleher and McCann, 1976); we take $\Delta \rho = 0.06$ g/cm$^3$ in our calculation. From the above values we have upward force per unit width, $P_b \approx 4 \cdot 10^{15}$ dyne cm$^{-1}$ and the maximum uplift at the plate edge, $y(0) \approx 4$ km. On the profiles of $AA'$ and $BB'$ of Fig. 2, the correspond-

![Fig. 5. Topographic profiles of the New Hebrides island arc system across lines $AA'$ and $BB'$, of Fig. 2, perpendicular to the trend of the island arc. The dashed curves show the bottoms of the basin after sediment is removed while the solid curves above the dashed curves are the sediments. $y(0)$ is the maximum deflection at the plate edge ($x = 0$); $x_{m1}$ is the distance from the plate edge to the first relative minimum, and $y(x_{m1})$ is the deflection at $x_{m1}$.](image)

ing values of \( y(0) \), which is supposed to be measured from the undeflected position of the plate in Fig. 5, should be measured from the averaged depth of the ocean floor to the east and far away from the island arc system to the highest point on the Santo or Malekula island. The measured value of \( y(0) \) from these two profiles is about 3.5–5 km which is consistent with the predicted value. The depth of the basin corresponding to \( y(x_{m1}) \) given in equation (5) is calculated to be 0.28 km. In order to measure \( y(x_{m1}) \) from Fig. 5 we must correct for the thickness of sediment in the basin, which is not well known. Luyendyk et al. (1974) report more than 1 km of undisturbed sediment in the basin. Here we use 1.3 km for the maximum thickness of the sediments. The dashed curves in Fig. 5 represent the basin bottoms after the sediment is removed. The measured value of \( y(x_{m1}) \) is about 0.4–0.7 km which agrees with the calculated value within a factor of three. Geological processes such as erosion, sedimentation, isostasy, and volcanic activity affect the topography as well, so we consider this agreement satisfactory. Our model, however, does not explain the topographic high associated with the Maewo and Pentecost islands (Fig. 5). A more complicated plate model than used above is necessary to explain these features.

In principle, the above analysis provides an estimate of the stress level within the bending plate. However, as demonstrated by De Bremeecker (1977), the estimate of stress depends very much on the constitutive equation and the model used in the calculation and is not very meaningful.

In this analysis, the upward loading alone is enough to explain the observations. The effects of introducing a horizontal loading at the plate edge are (1) to shift the location of \( x_{m1} \) or the basin center toward the plate edge, (2) to increase the amount of deflection, and (3) to increase the stress level within the plate. The first effect can bring the theoretical value of \( x_{m1} \) closer to the observation.

In the islands of the western belt (Santo and Malekula, Fig. 1), stratified sediments dip eastward some with dip angles larger than 60°, while those in the islands of the eastern belt (Maewo and Pentecost) dip westward with dip angles of about 20–30° (Mitchell and Warden, 1971; Luyendyk et al., 1974). This can fit our model very well.

In the present model the overriding plate is bent upward, so its upper part is subject to east—west compression. Chung and Kanamori (1978) studied

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**Fig. 6.** Topographic profile of the D'Entrecasteaux ridge-fracture zone system across line \( CC' \) of Fig. 2.
the focal mechanisms of the larger earthquakes in the area. Shallow earthquakes occurring within the overriding plate are characterized by high angle thrust faulting with horizontal compressional stress axes trending about east—west. This is probably due to the upward bending as well as the horizontal compression from the Australian (or Indian) plate. The seismic activity near the center of the basin is low. The events which occurred there are small so that no well-constrained mechanisms could be determined. Nevertheless, available first-motion data show no indication of normal faultings near the basin center. This finding together with Luyendyk et al. (1974), suggest that no active and recent sea floor spreading is taking place in the basin between 14\(^\circ\) and 16\(^\circ\)S. Our model requires no extensional mechanism to form the interarc basin. In addition, our model explains why the well developed basin occurs only at the extension of the ridges and not elsewhere in the arc.

CONCLUSIONS

In this study we have investigated several tectonic features associated with the subduction of aseismic ridges in the New Hebrides island arc. Our conclusions are:

1. The discontinuity of the New Hebrides Trench can be interpreted as a consequence of subduction of the aseismic ridges of the D'Entrecasteaux zone and the subsequent uplift of the trench-floor and plate edge due to the buoyant force associated with the low density ridges.

2. The islands of Santo and Malekula in the New Hebrides are probably formed by the uplifted outer arc or the mid-slope basement high, which is located between the trench and the frontal arc. Thus the two islands deviate from the general trend of the New Hebrides island arc.

3. The back-arc basin in the New Hebrides island arc system may have been formed as a result of subsidence of the overriding plate due to a vertical upward loading at the plate edge. This interpretation is consistent with geological observations.

4. The uplift at the plate edge and the subsidence behind the uplifted area can be modeled by a deflection of a thin elastic half plate overlying a fluid quarter space with a vertical upward loading at the plate edge. With reasonable values of the magnitude of the upward loading, the computed values of the maximum uplift at the plate edge, the distance between the basin center and the plate edge, and the depth of the basin agree with the observed values reasonably well.

Although the details of the tectonic features produced in the process of aseismic ridge subduction may be a function of convergence rate, mechanical properties of the overriding plate (continent or island arc), and the time duration from the beginning of the process (or the stage of plate interaction process), the phenomena investigated in this paper may be very general.
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