3. Tectonic Problems of Subduction Zones

RECENT TECTONICS ASSOCIATED WITH THE SUBDUCTION / COLLISION OF THE D'ENTRECASTEAUX ZONE IN THE CENTRAL NEW HEBRIDES

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ABSTRACT

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Recent data from the New Hebrides (Vanuatu) subduction zone have enabled a morphological description and tectonic interpretation to be made of the complex plate contact between the d'Entrecasteaux Zone and the West Torres Massif on the one hand, and the old Santo–Malekula block on the other. Between latitudes 14°30′ and 17°S no trench occurs owing to the presence of the old rigid Santo–Malekula block at the edge of the overriding plate. The outer limit of plate contact describes a tortuous line westward of the projected line of the adjacent trenches. This may be explained by local collision of the North d'Entrecasteaux Ridge, and the South d'Entrecasteaux Chain with the west flank of Santo. This collision is compensated for by lateral plastic flow of the inner wall downward into the small basins which separate the d'Entrecasteaux Zone ridges, a process which may be explained by plastic plane strain theory. The westward jump of the plate contact zone here may result from pieces of the North d'Entrecasteaux Ridge being cut off and incorporated in the frontal arc. A review of the geological evidence and the shallow seismicity of the central New Hebrides shows that radiant horizontal stresses resulting from this collision have generated a specific compressive stress regime which influences the whole width of the arc. The resultant deformation shows a close resemblance to a theoretical strain field produced by the indentation of a long, narrow, plastic body (cf. the arc) by a rigid, flat die (cf. the d'Entrecasteaux Zone). Thus the central New Hebrides block is both elastically bent under vertical stresses and plastically deformed and pushed eastwards by horizontal stresses. Recent uplift of the eastern chain can be explained by this model. In contrast, at the rear of both the northern and southern sections of the arc the tectonics are mainly controlled by a tensional stress regime. The strong plate interaction due to the subduction of the d'Entrecasteaux Zone is thought to be responsible for the compressive stress regime across the central New Hebrides. The same cause can be invoked to account for the tensional stress regime observed in north and south New Hebrides rear-arc regions.

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INTRODUCTION

The islands of the New Hebrides (Vanuatu) form a west facing intra-oceanic arc associated with the underthrusting of the Indo-Australian Plate beneath the North Fiji Basin. Unlike most peri-Pacific arcs, this subduction zone dips towards the Pacific. A typical arc-trench system is developed in the northern and southern parts of the arc, but the central area, between the Banks Islands and Efate (Fig. 1) has the following unusual features:
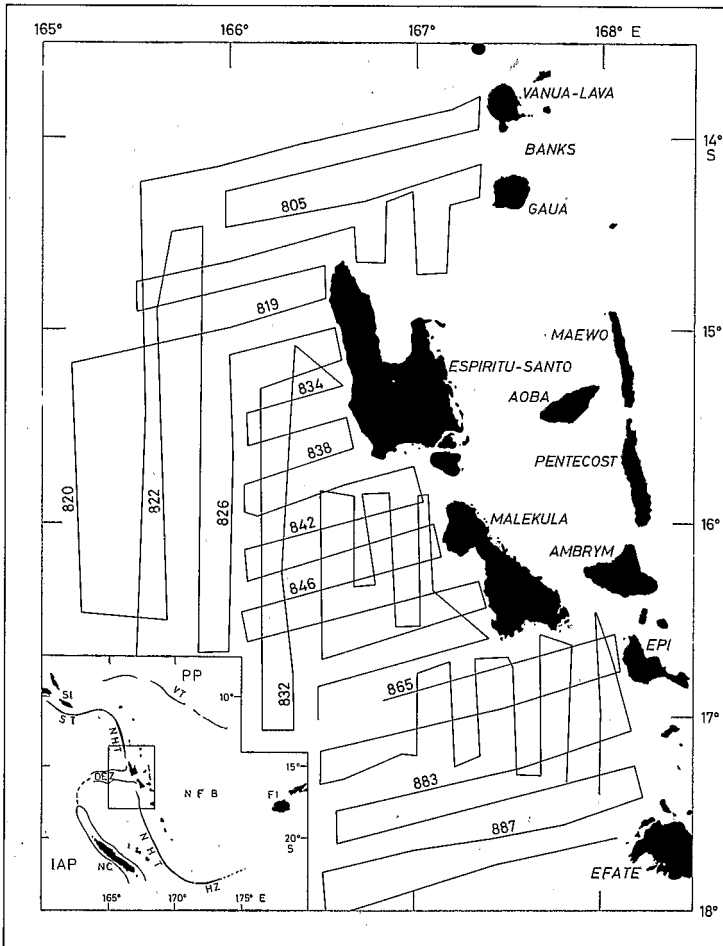

Fig. 1. Location of EVA VIII cruise tracks. Lower inset: regional setting. SI = Solomon Islands; ST = Solomon Trench; VT = Vitiaz Trench; PP = Pacific Plate; NHT = New Hebrides Trench; DEZ = D'Entrecasteaux Zone; NC = New Caledonia; IAP = Indo-Australian Plate; HZ = Hunter Zone; NFB = North Fiji Basin; FI = Fiji Islands.

(1) There is no trench between latitudes 14°30′ and 17°05′S (Karig and Mammerickx, 1972).

(2) The conventional arc pattern of a single island chain is replaced between latitudes 13°S and 17°S by a complex of three parallel chains (Mitchell and Warden, 1971). The islands of Espiritu Santo and Malekula form a western chain consisting of Oligocene–Miocene calc-alkaline lava flows and volcaniclastics. The islands of the eastern chain, Pentecost and Maewo, are of similar origin, but date from the Miocene–Pliocene. All those islands raised reef terraces that indicate several hundred metres of Pliocene–Quaternary uplift (Mallick, 1975; Jouannic et al., 1980; Taylor et al., 1980). These two chains are separated by the Aoba Basin which contains several thousand metres of Miocene–Pliocene sediments (Ravenne et al., 1977; Carney and MacFarlane, 1980, 1982; Katz, 1981, 1982) and a north–south chain of active volcanic islands (Gaua, Aoba, and Ambrym). Active volcanism in these islands is alkaline in contrast to the tholeiitic and calc-alkaline volcanism of the other volcanoes of the arc (Colley and Warden, 1974; Gorton, 1974, 1977; Roca, 1978).

(3) Microtectonic features of the active volcanic arc, such as the distribution of flank craters and the patterns of dykes, show that the maximum direction of compressive stress changes from northeast to southeast between Gaua and Epi (Roca, 1978).

(4) Isacks et al. (1981) demonstrated that thrust-type focal mechanism solutions for earthquakes along the whole arc have an average direction of slip of N76°E, a similar direction to that suggested by earlier work of Pascal et al. (1978). However, they show that there is significant variation from this value between 14° and 16°S.

(5) The maximum depth determined for epicentres of reliable intermediate earthquakes beneath the central region of the arc is only 170 km, with the greatest concentration occurring around 125 km (Louat et al., 1982; Marthelot, 1983). In the northern and southern regions of the arc, this depth increases and can reach 300 km.

These anomalous features of the central section of the New Hebrides arc have been related by several authors to the effects of the subduction of the d'Entrecasteaux Zone (Maillet et al., 1983), an area of high relief that extends westwards from Espiritu Santo (DEZ, Fig. 1) (Pascal, 1974; Choudhury et al., 1975; Ravenne et al., 1977; Pascal et al., 1978; Chung and Kanamori, 1978a, 1978b; Roca, 1978; Jouannic et al., 1980; Taylor et al., 1980; Isacks et al., 1981; Jouannic et al., 1982; Marthelot, 1983). However, Daniel and Katz (1981) concluded that the absence of a trench in the central part of the New Hebrides arc is due to the existence of old resistant insular blocks at the leading edge of the overriding plate at the start of subduction.

The subduction of the d'Entrecasteaux Zone and the presence of the resistant Santo–Malekula block in the position normally occupied by the arc-trench gap are, therefore, considered to be the two major factors influencing the deformation of the plate contact zone in the central New Hebrides.

In this paper we first describe the nature of the underthrusting plate. Next we

trace the boundary between the two converging plates, and interpret the deformation of the leading edge of the overriding plate. Finally we analyse the stress and strain patterns across the entire arc.

MATERIAL STUDIED

This study is based on bathymetric and magnetic data collected during October 1981 on a cruise of the R.V. "Coriolis" (Fig. 1). The objectives of this ORSTOM-organised cruise were to examine the relationship between the frontal arc, comprising the islands of Espiritu Santo and Malekula, the deep-water re-entrant northwest of Efate, and the d'Entrecasteaux Zone. Observations were made with a 12 kHz echo sounder and a proton precession magnetometer. Navigation was by fully automatic satellite system with dead reckoning positions calculated from ship's heading and speed every minute between fixes and corrected with the next fix. Bathymetric and magnetic maps are shown in Figs. 3 and 4, respectively.

These data have been augmented by seismic reflection profiles of GULF and MOBIL as well as some of the seismic reflection profiles obtained during the October 1982 GEOVAN II cruise of the R.V. "Coriolis", which was organised jointly by CCOP/SOPAC and ORSTOM (Fig. 2).
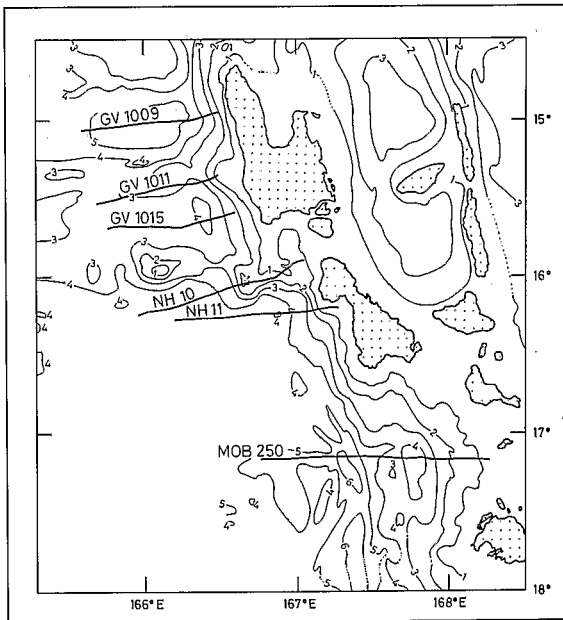

Fig. 2. Location of seismic lines cited in text. *MOB* = Mobil 1972 cruise; *NH* = Gulf 1973 cruise; *GV* = Geovan II (ORSTOM–CCOP/SOPAC) cruise 1982. Isobaths in km.
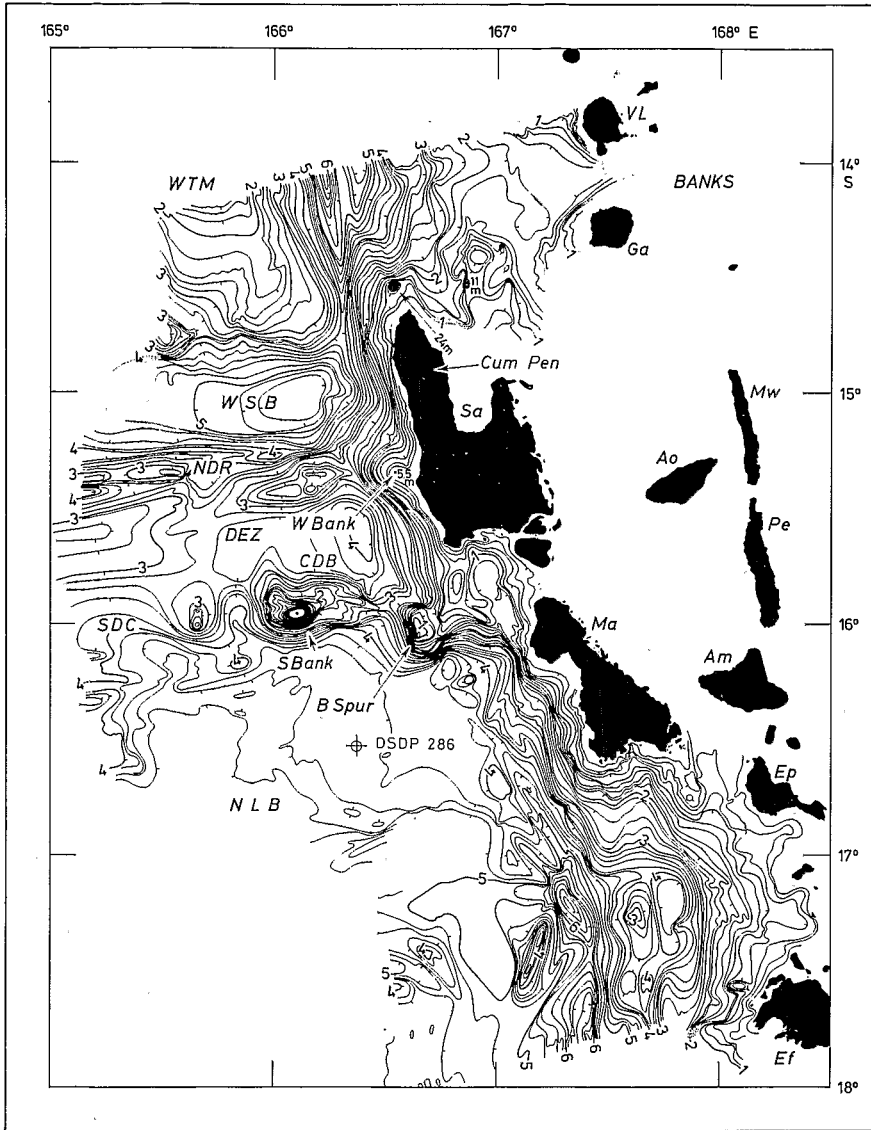
Fig. 3. Bathymetric map of the eastern d'Entrecasteaux Zone and surrounding areas. Isobaths are at 0.2 km intervals. Soundings in meters are specified on some highs. *WTM* = West Torres Massif; *WSB* = West Santo Basin; *Cum Pen* = Cumberland Peninsula; *NDR* = North d'Entrecasteaux Ridge; *DEZ* = d'Entrecasteaux Zone; *WBank* = Wousi Bank; *CDB* = Central d'Entrecasteaux Basin; *SDC* = South d'Entrecasteaux Chain; *SBank* = Sabine Bank; *BSpur* = Bougainville Spur; *NLB* = North Loyalty Basin. Names of the islands are indicated in Fig. 1.
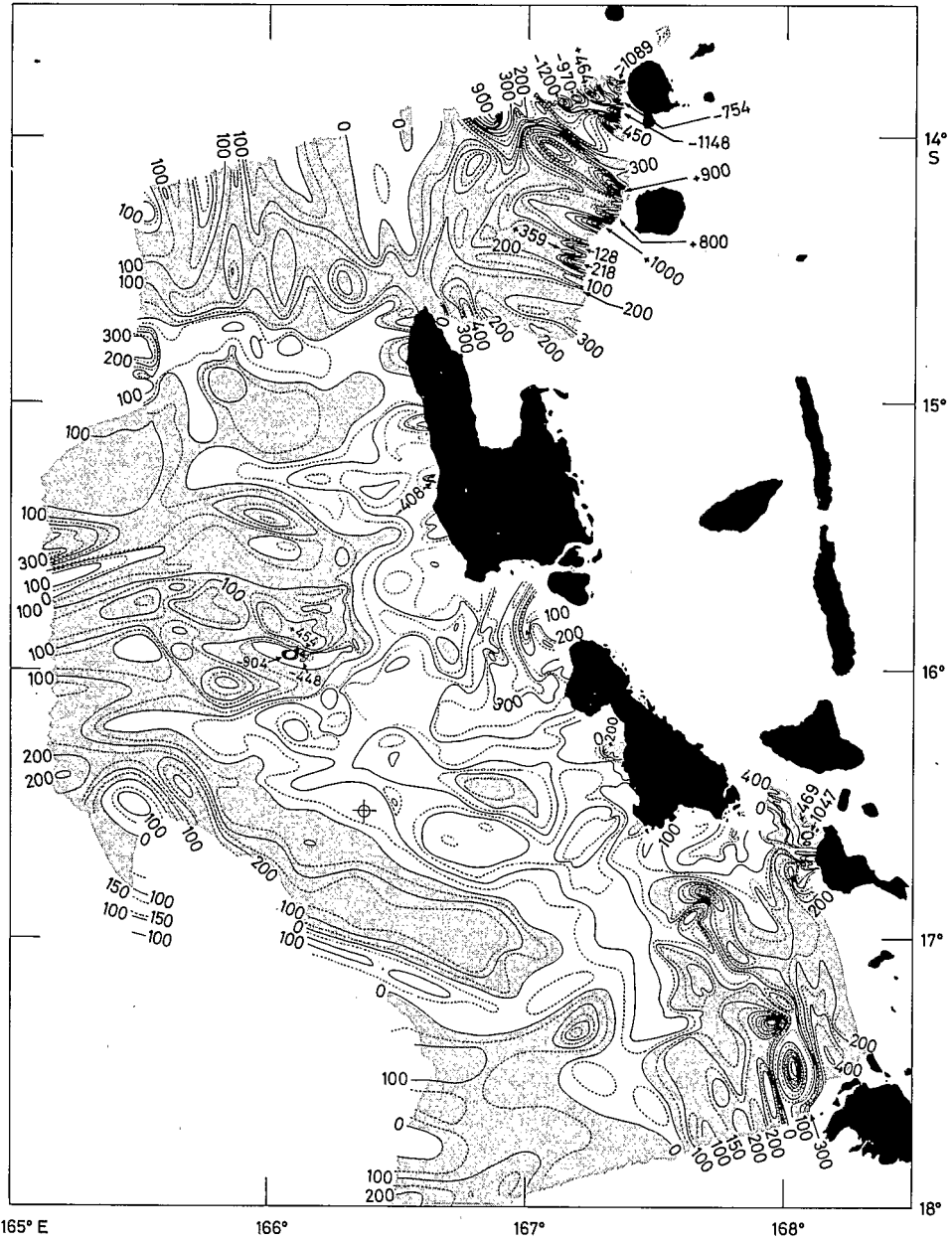
Fig. 4. Magnetic map—magnetic anomalies, obtained from total magnetic field after removing the IGRF 1975, are contoured at 50-gamma intervals. Shaded areas are positive anomalies and white areas are negative anomalies. The crossed circle is the location of the DSDP drilling site 286.

DATA ANALYSIS

The nature of the subducting plate

The major topographical features that occur from south to north along the margin of the subducting plate are (Fig. 3), the North Loyalty Basin (NLB), d'Entrecasteaux Zone (DEZ), West Santo Basin (WSB), and West Torres Massif (WTM).

(1) *North Loyalty Basin.* The irregular floor of the southern part of this basin becomes smoother north of latitude 17°S. The basin is underlain by an oceanic crust 12 km thick (Ibrahim et al., 1980; Pontoise et al., 1980) and at least 47 Ma old. It contains 650 m of sediments at DSDP site 286 (Fig. 3) (Andrews et al., 1975). Water depths decrease from 4900 m in the central part of the basin to 4300 m at the plate boundary facing Malekula.
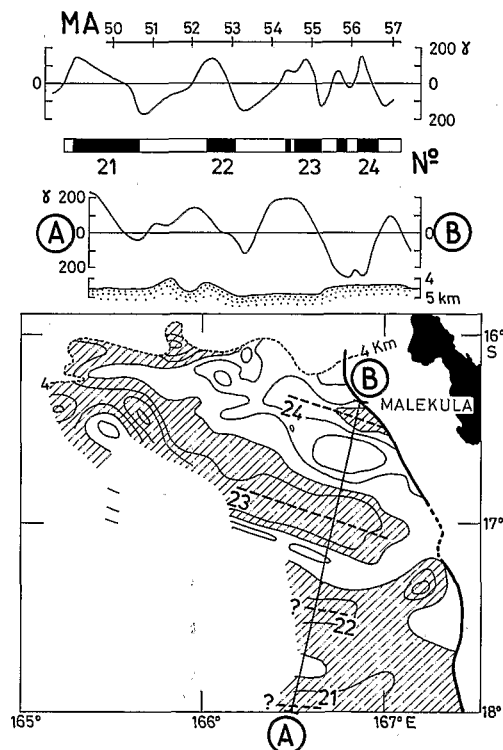

Fig. 5. Magnetic interpretation of the northeastern part North Loyalty Basin. *Top*: Theoretical anomaly after Lapouille (1982) calculated for a strike direction of N50°E; half spreading rate of the 2.2 cm/yr magnetization of 0.0048 emu/cm³ with a magnetised sheet thickness of 0.5 km located at a 5 km depth. *Center*: Residual magnetic anomaly cross-section plotted against the bathymetric section. *Bottom*: Magnetic map—ruled areas are positive magnetic anomalies. Dashed lines indicate magnetic anomaly trend with the corresponding number.

Magnetic anomalies over the basin are shown in Fig. 4. The lineations generally trend N110°, and have amplitude and wavelength characteristics attributable to geomagnetic reversals correlated from south to north with anomalies 21 to 24 (Fig. 5). Weissel et al. (1977, 1982), and Lapouille (1982) have described a similar pattern from the southeastern part of the basin where the anomalies were dated on the magnetic reversal time scale at 50–56 Ma, giving an inferred half spreading rate of approximately 2 cm/yr. The NE–SW lineation trend proposed by Lapouille (1982) and Weissel et al. (1982) is substantially different from that we observed. However, the contoured magnetic data (Fig. 4) suggest that small fracture zone offsets could exist, trending approximately N–S and occuring from the NW to the SE. Therefore, the true direction of the magnetic lineations could be E–W, a direction close to that defined by Watts et al. (1977) for the western part of the adjacent South Fiji Basin.

(2) *The d'Entrecasteaux Zone.* This is a continuous arcuate structure up to 100 km wide that rises sharply from the North Loyalty Basin and extends from the New Caledonia–Loyalty Islands platform eastwards to the central New Hebrides (Fig. 1). Its western and central portions have been mapped by Jouannic (1975) and studied by Daniel et al. (1977) and Maillet et al. (1983).

The western part of the eastern d'Entrecasteaux Zone (165°–165°40′E) is a slightly asymmetrical massive structure with mean depth of about 3500 m, generally bounded on its north by a scarp (Fig. 6 and Fig. 7 E820). Dredge hauls from this scarp have recovered volcanic tuff and mid-oceanic ridge type basalt of early Eocene age (Maillet et al., 1983). MOR type basalt from a steep-sided narrow ridge along the north side of the scarp has been dated at 37 Ma. Maillet et al. (1983) have interpreted these structures as pieces of oceanic crust uplifted and tilted during a middle Miocene to Quaternary tensional phase, after an upper Eocene compressional phase, the ultimate manifestation of which was the obduction of ultramafics in New Caledonia (Collot et al., 1982).

East of longitude 165°40′ the massive structure of the zone is gradually replaced by a small basin, the Central d'Entrecasteaux Basin (CDB) covered by 1 sec (two-way time) of sediments (Luyendyk et al., 1974) which gives rise to a relatively smooth sea floor, gradually deepening to 4000 m near the west flank of Santo. This basin is flanked by two ridge systems. To the north lies the North d'Entrecasteaux Ridge (NDR) (Figs. 3 and 7, E826 and E832) which trends N76° to the Wousi Bank, a morphological extension of Santo. The shallowest depths of the ridge are 2350–2100 m, and it is considered to be a continuation of the structural unit described to the west by Maillet et al. (1983). South of the basin lies a seamount chain, the South d'Entrecasteaux Chain (SDC), which includes the Sabine Bank, which rises to only 7 m below sea level. A small 2300 m deep seamount lies west of the Sabine Bank. The Sabine Bank is a strongly magnetic conical feature superimposed on a broader E–W trending ridge, the whole structure being bounded by a scarp that falls 4000 m to the floor of the North Loyalty Basin (Fig. 6). The seamounts, interpreted as volcanic edifices on the basis of their morphology, are possibly the remnants of an old island

arc (Lapouille and Dugas, 1975) that could have been the source of the upper Eocene andesitic debris encountered at DSDP site 286 (Maillet et al., 1983). To the east of the seamounts lies the Bougainville Spur, which is morphologically connected to the southwestern extension of Espiritu Santo. It is well delineated and has a long axis parallel to the general trend of the New Hebrides. Daniel and Katz (1981) have
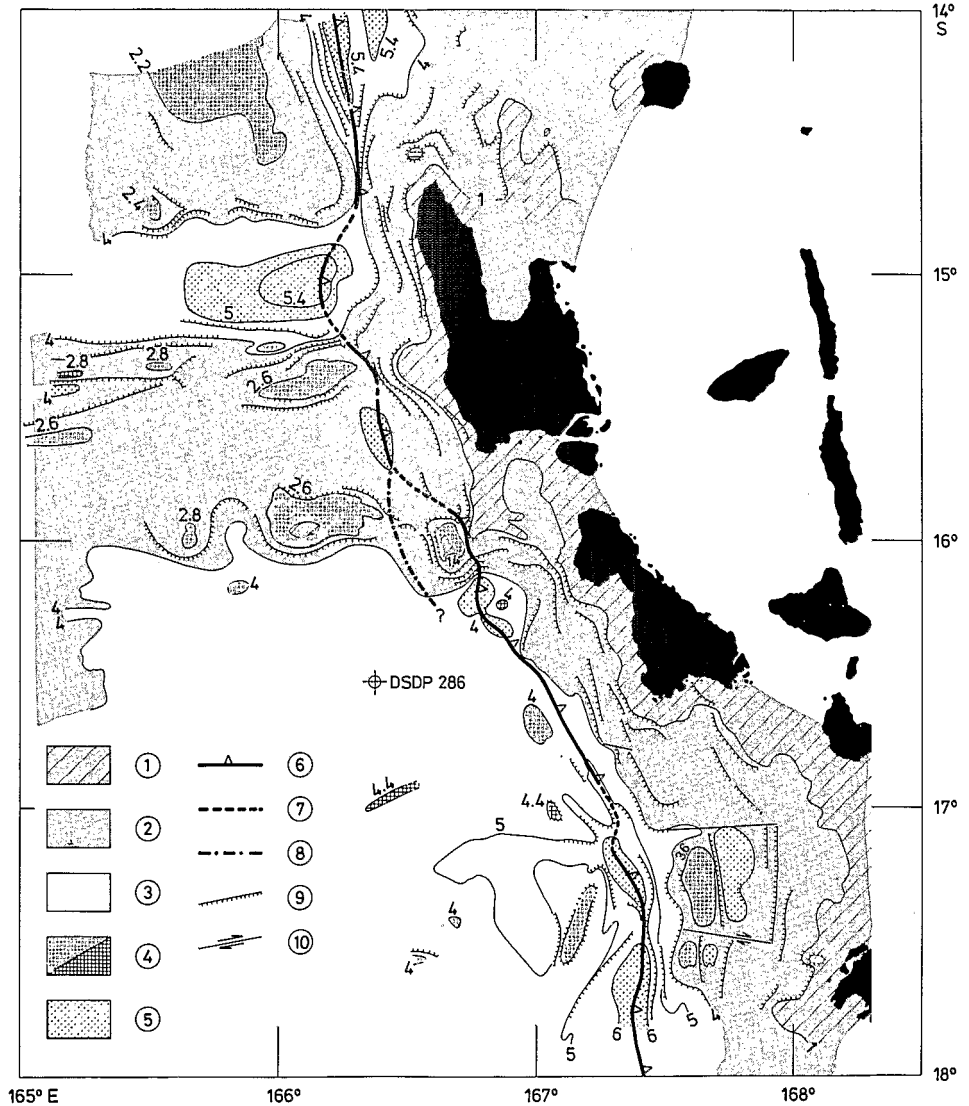

Fig. 6. Diagrammatic interpretation of the bathymetric map. Isobaths are in km. *1* = depths less than 1 km; *2* = depths less than 4 km; *3* = depths greater than 4 km; *4* = relative topographic highs; *5* = relative topographic lows; *6* = observed outer limit of the plate contact zone; *7* = conjectural outer limit of the plate contact zone; *8* = expected evolution of the outer limit of the plate contact zone; *9* = major scarps; *10* = apparent sense of offset defined from bathymetry.
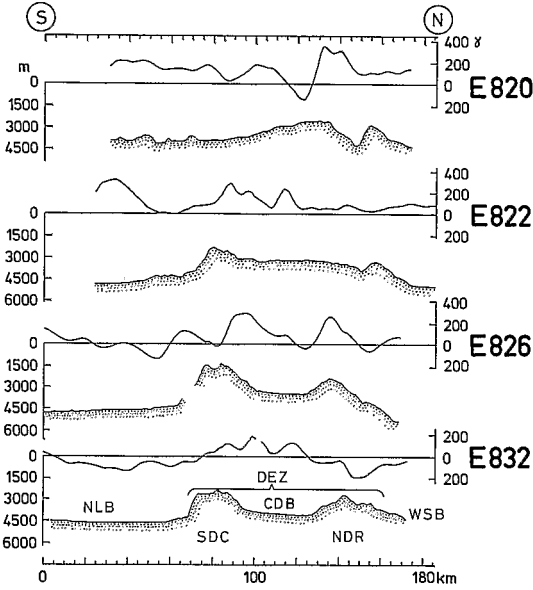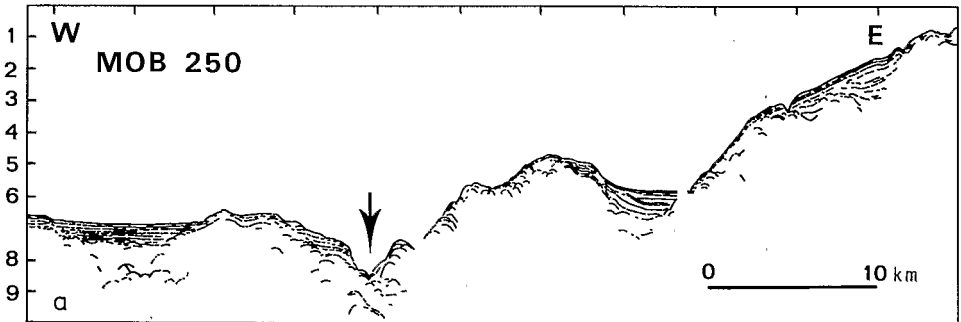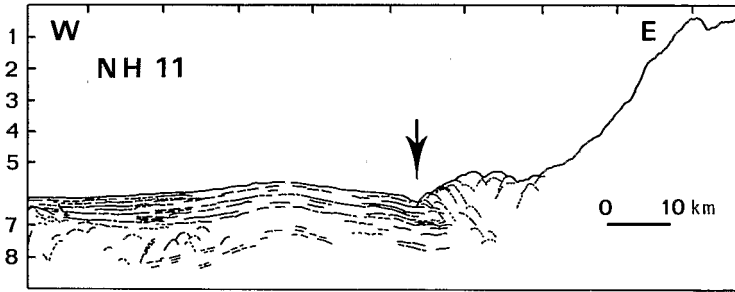
Fig. 7. Bathymetric and magnetic profiles across the d'Entrecasteaux Zone. Location of profiles is shown in Fig. 1. Abreviations as with Fig. 3.
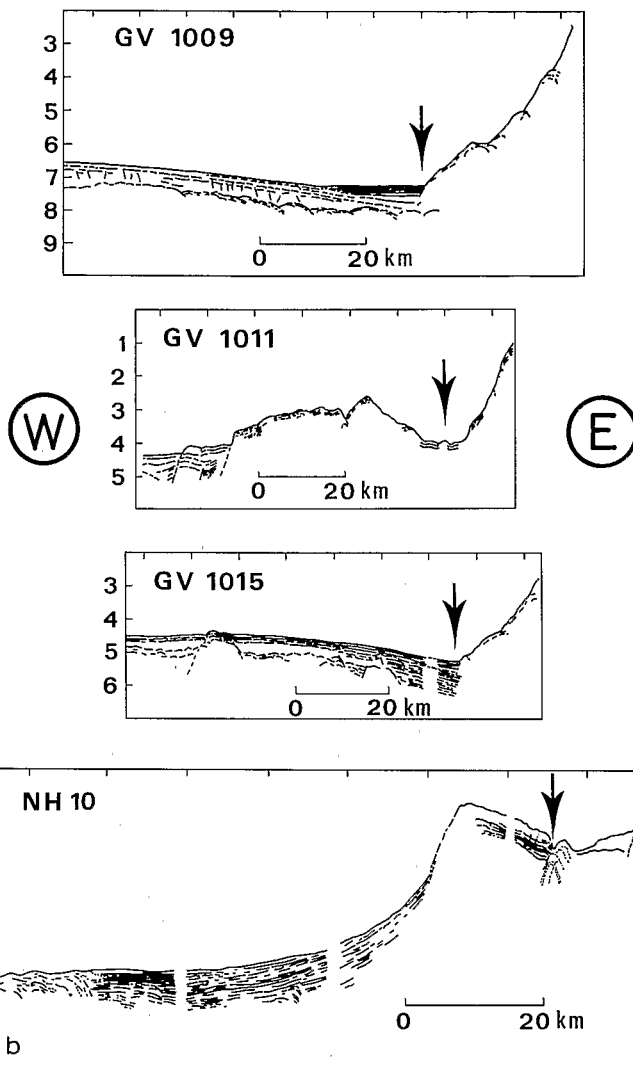
Fig. 8. Seismic reflection profiles across the plate contact zone. See Fig. 2 for locations. Vertical scale is in sec (two way time). Arrows emphasise the outer limit of the plate contact zone.

suggested that this could be an uplifted piece of oceanic crust.

The magnetic anomalies of the eastern d'Entrecasteaux Zone are generally a few hundred gammas in amplitude. They are consistent with the trend of the ridge structures and with them being of oceanic or volcanic origin.

We interpret the eastern d'Entrecasteaux Zone as an old plate boundary that originated in the pre-Miocene as a south-dipping subduction zone.

(3) *West Santo Basin*. The floor of this 25 km wide basin trends smoothly

eastward, deepening gradually to 5400 m at the toe of Espiritu Santo. Although this water depth is typical of oceanic crust, the seismic reflection profile 104 in Daniel et al. (1977) shows that the basin contains more than 1.2 sec (two way time) of sediment. The roughly N–S asymmetrical shape of the basin displayed on this seismic profile suggests that it may be a former E–W trench-like feature.

(4) *West Torres Massif.* This feature is about 200 km long and 150 km wide. The minimum water depth is about 1100 m (Jouannic, 1975), and Dupont and Recy (1981) speculated that it is an oceanic plateau with an intermediate crust.

The nature of the plate boundary

An intra-oceanic convergent plate boundary generally follows the deep axis of the trench and is easily defined. However, in the central New Hebrides a trench does not exist and the surface trace of the plate boundary is complicated by subduction complexities produced by the interaction of the two plates. In this region we have found it useful to define the plate boundary at *the outer limit of the plate contact zone* (Fig. 6), where the area composed exclusively of lithologies originating on the downgoing plate abuts the zone of subduction complex. This has been interpreted from seismic reflection (Fig. 8) and bathymetric profiles (Fig. 9).

West of Efate and Malekula, the plate boundary follows the axis of the well-developed trench as far north as 17°05′S (Fig. 3; MOBIL 250, Fig. 8a; E887, E883, Fig. 9). No trench is recognised on line E865 (Fig. 9) and the plate boundary has been drawn along the axis of the shallow depression at the foot of the insular slope.

West of Malekula, there is an abrupt transition from the relatively undisturbed floor of the North Loyalty Basin to the disturbed toe of the insular slope (Line E846, Fig. 9). Line NH 11 (Fig. 8a) shows underthrusting of the North Loyalty Basin beneath the insular slope (Daniel and Katz, 1981). In this region the plate boundary has been drawn along the junction of the smooth abyssal plain and the rugged insular wall.

The change in strike of the plate boundary from northerly, west of Efate, to north-northwesterly, west of Malekula occurs at point corresponding to the disappearance of the trench at latitude 17°05′, and may reflect the influence of the Malekula structure.

East of the d'Entrecasteaux Zone, the plate contact zone is complex (Fig. 8b; see also Luyendyk et al., 1974; Daniel and Katz, 1981). No deep morphological trench has developed, instead the western flank of the Santo block is abutted by a deep basin, the Central d'Entrecasteaux Basin (GEOV 1015, Fig. 8b; E838, Fig. 9), bounded by two topographical highs, the North d'Entrecastraux Ridge and the South d'Entrecasteaux Chain (GEOV 1011, NH10, Fig. 8b; E834–E842, Fig. 9).

Seismic profile GEOV 1011 (Fig. 8b) shows recently deformed sediments on the western part of the North d'Entrecasteaux Ridge. On the eastern flank of the ridge these sediments have either become totally chaotic or are absent. However, a small

accumulation of recently deformed sediments occurs in the depression between the North d'Entrecasteaux Ridge and the steep wall of the Wousi Bank (a feature contiguous with the Santo block). We have drawn the plate contact along this tectonically disturbed saddle (E834, Fig. 9; Fig. 6).

In the Central d'Entrecasteaux Basin there is evidence of underthrusting of the basin beneath the Santo block (GEOV 1015, Fig. 8b). Despite its shallower depth, this section is comparable to GEOV 1009 (Fig. 8b) from the West Santo Basin described below. The plate boundary has been drawn at the base of the insular slope (Fig. 8b; E838, Fig. 9).
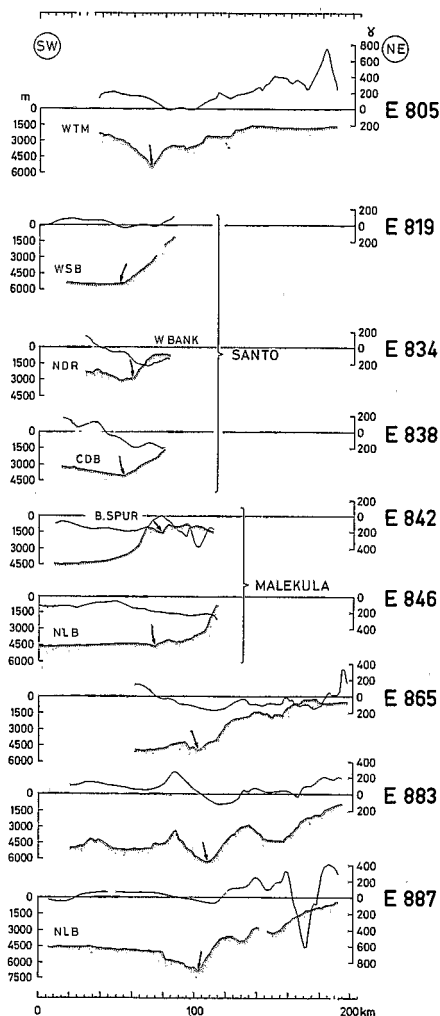

Fig. 9. Bathymetric and magnetic profiles across the plate contact zone. Location is shown in Fig. 1. Arrows emphasise the outer limit of the plate contact zone. Abreviations as with Fig. 3.

The South d'Entrecasteaux Chain abuts the Bougainville Spur which forms a topographic elevation bounded on three sides by very steep walls (Fig. 3). The west wall of the spur shows upthrust sediments on seismic section NH10 (Fig. 8b). The spur is capped by sediments similar in character to those on the oceanic plate to the west, but which are markedly different to the disturbed sediments on the rugged acoustic basement that forms the southwest extension of Santo. We have followed the proposal of Daniel and Katz (1981) and have drawn the plate boundary along the small depression east of the Bougainville Spur, since this divides the two areas of contrasting sediment character.

Immediately north of Espiritu Santo, the well-developed trench is split into two by a submarine prolongation of the Cumberland Peninsula of Santo (Fig. 3). The plate boundary has been drawn along the axis of the western trench, which is 6300 m deep, the eastern branch being 5600 m deep. The western trench extends northwards as a single progressively deepening feature, reaching a maximum depth of 9174 m (Mammerickx et al., 1974; Faleyev et al., 1977).

West of the Cumberland Peninsula of Santo the trench gradually shallows southwards to a point where its floor is only 4070 m deep, a depth similar to that of the North Loyalty Basin. However, a topographical trench still exists as a saddle between the West Torres Massif and Espiritu Santo (E805, Fig. 9). South of the West Torres Massif this depression opens onto the West Santo Basin and the steep sided trench disappears (Fig. 6). However, seismic profile GEOV 1009 (Fig. 8b) clearly indicates a landward deepening of the superficial structures consistent with the underthrusting of the West Santo Basin beneath the Santo block. The plate contact has been drawn along the toe of the insular slope (E819, Fig. 9).

INTERPRETATION

In conventional intra-oceanic zones of plate convergence deformation of the overriding lithospheric plate is quite often extensional (Nakamura and Uyeda, 1980), and large compressive stresses are not usually transmitted across the arc. However, when a ridge, massif, or continental protrusion occurs on the underthrusting plate and the edge of the overriding plate is occupied by an isolated, old, resistant insular block, then subduction would be expected to generate horizontal compressive stresses. So long as these do not become strong enough to halt subduction, then subduction would continue accompanied by local compressive deformation affecting mostly the overriding plate, but also the front of the underthrusting plate. Depending on the size and nature of the colliding object, the area affected by compressive deformation could extend across most of the arc (Dickinson and Seely, 1979). We will, therefore, first consider the interpretation of the deformation of the zone of plate contact, and then consider the deformation of the whole width of the arc.

Contact zone deformation

Density heterogeneity within a plate may reduce its flexure (Moretti, 1983), and the effects of this would be further emphasised by ridge topography. Therefore, the subduction of a ridge would theoretically be expected to cause both a shift of the plate boundary towards the arc and a shallowing of the trench (Vogt et al., 1976). These effects are associated with the subduction of the Louisville Ridge beneath the Tonga arc (Dupont, 1979, 1982), but are not seen in the central New Hebrides, where the plate boundary is displaced westward, away from the arc, and no trench is developed. These effects seem to result from the protrusion of the Santo–Malekula block at the leading edge of the overriding plate. However, the tortuous trace of the plate boundary may reflect the effects of collision of the North d'Entrecasteaux Ridge and South d'Entrecasteaux Chain with the Santo block.

At most convergent plate boundaries the thickness of the plates generally determines that flexure will occur along a hinge with a smooth and modestly curved plan view (Dickinson and Seely, 1979). Even when a subduction zone is located along the jagged edge of a large crustal block, the trend of the trench is still quite smooth. A linear trench of this type has developed in the New Hebrides region north of 14°15′S and south of 17°05′S. At 17°25′S, there is a local interruption of the trench, east of a 1.6 km high, N20°-trending seamount, and at a point coinciding with a sharp interruption of the N–S trending magnetic anomalies (Fig. 4). Here, the trench floor rises to 5750 m. This feature may represent the incipient effects of the seamount on the downgoing plate impinging on the overriding plate, resulting in local deformation as well as offset of the adjacent fore-arc (Fig. 6). The seamount may also have suffered deformation, for it is flanked by features that may be active faults (profiles E883 and E887, Fig. 9), and Chatelain et al. (1983), who studied the earthquake of 25 July 1981, noted a great density of aftershocks that were propagated in this area.

The deformed trace of the plate boundary commences at the northern end of the southern New Hebrides Trench, a point that lies west of a major along-strike transition in the overall structure of the island arc. The relative gap in the distribution of shallow moderate-sized earthquakes around latitude 17°S (Fig. 11) suggests a major along-strike change in the stress regime in this area (Isacks et al., 1981), probably owing to a decoupling of plates in a transitional zone.

However, the outstanding feature of this zone is the drastic alteration of the frontal arc morphology between Efate and Malekula, where the emergent island of Malekula abruptly replaces a deep water re-entrant that reaches depths of 4200 m. This deep-water area consists of a small flat-bottomed trough separated from a trench by a narrow ridge (Fig. 6; E883, Fig. 9). These features appear to be the results of conventional fore-arc sedimentary processes in a segment of oceanic crust trapped between trench and arc (MOBIL 250, Fig. 8a). Thus the re-entrant may be regarded as a fore-arc basin, and the ridge as a subduction complex. However, in the

absence of more data, the ridge could equally be interpreted as an uplifted piece of basement bounded by normal faults.

South of 17°S, the bending of the downgoing plate is controlled by normal faulting. North of this latitude, there is evidence of crustal updoming (NH 11, Fig. 8a) which increases in intensity northwards, until part of the subducting plate is actually upthrust and deformed to form the Bougainville Spur (Daniel and Katz, 1981). This suggests that the stress field of the downgoing plate, which is mainly tensional south of 17°S, becomes compressional west of Malekula.

Stresses reach high values at the position of the upthrust Bougainville Spur. This stress pattern could be explained by an extremely localised strong coupling of the plates beneath the submarine extension of Santo probably caused by the previous subduction of the South d'Entrecasteaux Chain.

The outer limit of the plate contact zone in front of Santo (Fig. 6) is tortuous and located westward of the projected axis of the trenches to the north and south. This boundary is poorly controlled by data and unrecognised discontinuities, such as strike–slip faults might exist connecting the displaced portions of the plate boundary.

The lack of a deep trench north of latitude 15°15′ cannot be attributed to the effects of subduction of the d'Entrecasteaux Zone, since this is the present northern limit of the zone, and it is being subducted along a N76°-trending regional slip vector.

The incipient contact of the West Torres Massif with the western edge of Santo is responsible for producing a trench-like morphology with a floor at normal oceanic depth. The approach of the West Torres Massif towards the arc, combined with the bending of the plate has given rise to its steep eastern flank controlled by active faults, as is confirmed by the distribution of shallow moderate sized earthquakes (Fig. 11). The steep scarps affecting the northwestern flank of Santo reflect the same process. However, we think that the recent arrival of the West Torres Massif at the plate boundary is not basically responsible for the absence of a deep trench, but that this is due to the presence of the resistant 20–25 Ma old Santo block protruding at the leading edge of the overriding plate since the start of subduction (Daniel and Katz, 1981). The same explanation may be invoked to explain the development of only incipient trenches at the plate boundary next to the conventionally subducting West Santo and Central d'Entrecasteaux basins.

Thus, when subduction occurs along a crustal boundary bearing a relatively small and localised rigid protrusion, the necessary continuity of the flexure of the downgoing plate may dictate that the topographical expression of the flexure, i.e. the morphological trench, although developed either side of the protrusion, disappears as the zone of flexure passes underneath the protrusion, it being impossible for the zone of flexure to be diverted around the front of the protrusion.

However, the deformation of the slope of the Santo block and the resultant masking of the trench might be due in part to the seaward displacement of the lower insular slopes of Santo to compensate for compression in the zone of collision

between the North d'Entrecasteaux Ridge, the South d'Entrecasteaux Chain and the Santo block. This deformation may be explained by reference to a plane strain plastic flow model.

Since increasing temperature generates plastic deformation, the deformation of the crust in depth is likely to be controlled by creep. Thus, as the crust and lithosphere of the overriding plate are heated during subduction, they could behave plastically if indented or squashed by colliding objects borne on the downgoing plate, although plain strain plastic flow can only occur when the maximum shear stress reaches the limiting value of the yield stress for the material in question. The main properties of the plane strain plastic medium have been summarised by Tapponnier and Molnar (1976) and Molnar and Tapponnier (1977), and Fig. 10A shows the geometry of slip lines that would result from the simple case of the indentation of a semi-infinite rigid-plastic body by a flat rigid die.

In the central New Hebrides arc, the frontal arc west of Santo is deformed by three ridge-like features (SDC, NDR, WTM in Fig. 10B) that may be compared to three rigid dies separated by two deep basins. We can apply the model illustrated in Fig. 10A to predict the deformation pattern of the overriding plate edge in this region by combining the slip lines that would theoretically be created by each of the dies (Fig. 10B). This indicates that the collision and indentation of the dies may be compensated for laterally by plastic flow at the inner wall of the trench next to the intervening basins. The material thrust aside by the indenting front would appear to be preferentially extruded along left-lateral and right-lateral slip lines towards the deep basins, where subduction of superficial material would be more easily accomplished. Consequently, the outer limit of the plate contact zone would move progressively away from the arc westward across these small basins. Furthermore, the vertical deformation shown by the recent uplift of the arc (Mallick and Greenbaum, 1977; Taylor et al., 1980; Jouannic et al., 1980) probably facilitates the movement of material into the deep basins. We therefore speculate a horizontal component in the tectonics of insular slope (i.e. the inner wall) which contrasts with the vertical component that controls the tectonics of the top of the frontal arc. This speculative model of the tectonics of the inner wall is consistent with the overall fault pattern observed on Santo (Fig. 10B). Although most of these faults are said to be of Miocene age, Mallick and Greenbaum (1977) considered that they have probably been reactivated since this time, and evidence for recent reactivation is found in arcuate normal faults that downthrow westwards and control the west coast of Santo.

The difference in mechanical behaviour between the subducting ridges and the overriding block in this interpretation may be explained by the relative age differences between the d'Entrecasteaux ridges, which are of late Paleocene–early Eocene age (Maillet et al., 1983) and, therefore, might be assumed to be cooler and denser than the younger Santo block, dated as late Oligocene by Carney and Macfarlane (1982).

The superficial effects of the collision of the ridges of the d'Entrecasteaux Zone with the Santo block should not be confused with the gigantic phenomenon of continental collision. Scales and duration are so different that effects must also be different. The local earthquake pattern shows that the main component of the system is still subduction. However, the local compressive stresses resulting from the interaction between the irregularities of the d'Entrecasteaux Zone and the Santo–Malekula block have resulted in significant, though apparently superficial, collision-type deformation.

This combined subduction/collision process may allow the arc to grow by the accretion of scraped or obducted pieces of the downgoing plate onto the frontal arc. Thus the Bougainville Spur, which we locate west of the plate contact zone, appears to be in a position of being forced beneath the arc by subduction processes. As its
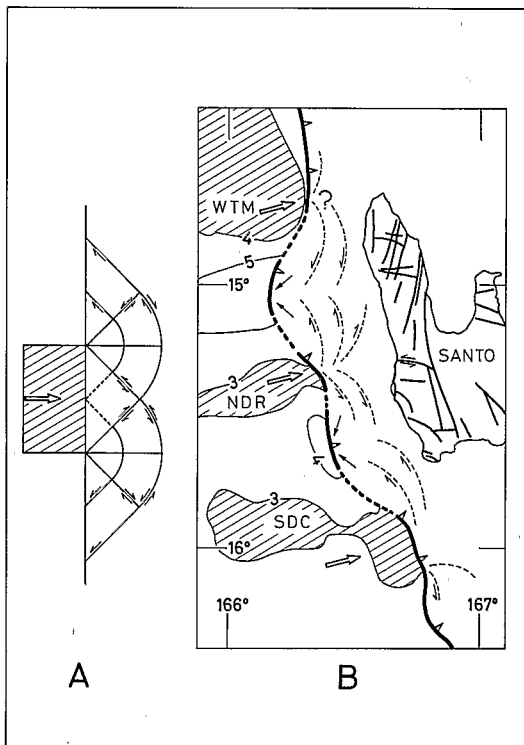

Fig. 10. A. Geometry of slip lines in the case of plate indentation of a semi-infinite rigid-plastic medium by a flat rigid die. Arrows indicate sense of shear along slip lines (after Taponnier and Molnar, 1976). B. Plate contact zone deformation West of Santo. Ruled areas mark the ridges (West Torres Massif, North d'Entrecasteaux Ridge, South d'Entrecasteaux Chain) identified as rigid dies. Light dashed lines with arrows indicate inferred slip lines with assumed sense of displacement. Solid lines on Santo show the observed fault pattern (after Carney and Macfarlane, 1982). Open arrows indicate the general direction of plate convergence. Broad line is the outer limit of the plate contact zone (see Fig. 6). Bathymetry in km.

strength is too great to permit its complete subduction, it seems inevitable that it will eventually be cut off from the downgoing plate. This will result in the outer limit of the plate contact zone jumping to the west of the spur, which will be left as a protruding part of the frontal arc (Fig. 6). The compressive stresses associated with this process could generate reversed faulting in the frontal arc. The Wousi Bank, which is located eastward of the outer limit of the plate contact zone could be a former outlier of the North d'Entrecasteaux Ridge that was incorporated onto the Santo block by this process at an earlier stage. This interpretation is supported by the occurrence of some reversed faults dipping southwest at 47°–60°, reported from the western part of central Santo by Mallick and Greenbaum (1977) and by a curved sinistral strike–slip, fault reported by Carney and Macfarlane (1982) (Fig. 10B). Though most of these faults are dated as Miocene, their reversed movement is probably more recent.

The plate boundary in the central New Hebrides presumably initially ran straight along the flank of Santo, and the westward shift of the plate contact zone is, therefore, probably partly accounted for by the effects of the subduction/collision of the d'Entrecasteaux Zone.

Stress and strain across the arc

The pattern of the stress field responsible for the present geological configuration of the central arc can be reconstructed from a study of geological structures developed during the Plio-Quaternary, and from the records of shallow seismicity collected over the last two decades. A review of these data has been undertaken to assess the influence of the collision/subduction process across the entire width of the arc.

(1) *Structural evidence:* The generally elongate shapes of the active Quaternary volcanoes of Aoba and Ambrym (Fig. 11), the occurrence and trend of aligned adventive craters, and the trend of dyke patterns indicate the direction of maximum horizontal compressive stresses across the arc. Microtectonic and fault pattern analysis indicates NE–SW to N–S radiating stress directions in the South Aoba area (Roca, 1978), converging on the point of collision of the South d'Entrecasteaux Chain (The Bougainville Spur) with the southwestern extension of Santo (Fig. 11). The Banks Islands, located north of the Aoba Basin, are roughly circular and poorly deformed; however, Roca (1978) and Katz and Daniel (1981) have noted NE–SW to N–S preferred directions of flank-crater distribution in these islands. This suggests a relative symmetry of the central New Hebrides stress field pattern with respect to the N76 direction of subduction of the d'Entrecasteaux Zone. The observed stress direction varies from S–N in the northern Aoba area, through NE–SW and E–W, to N–S in the Epi, Efate area, and this is interpreted as a consequence of the d'Entrecasteaux–Santo subduction/collision. The undeformed shape of the Banks volcanoes could suggest that either horizontal stresses are smaller or transmitted less
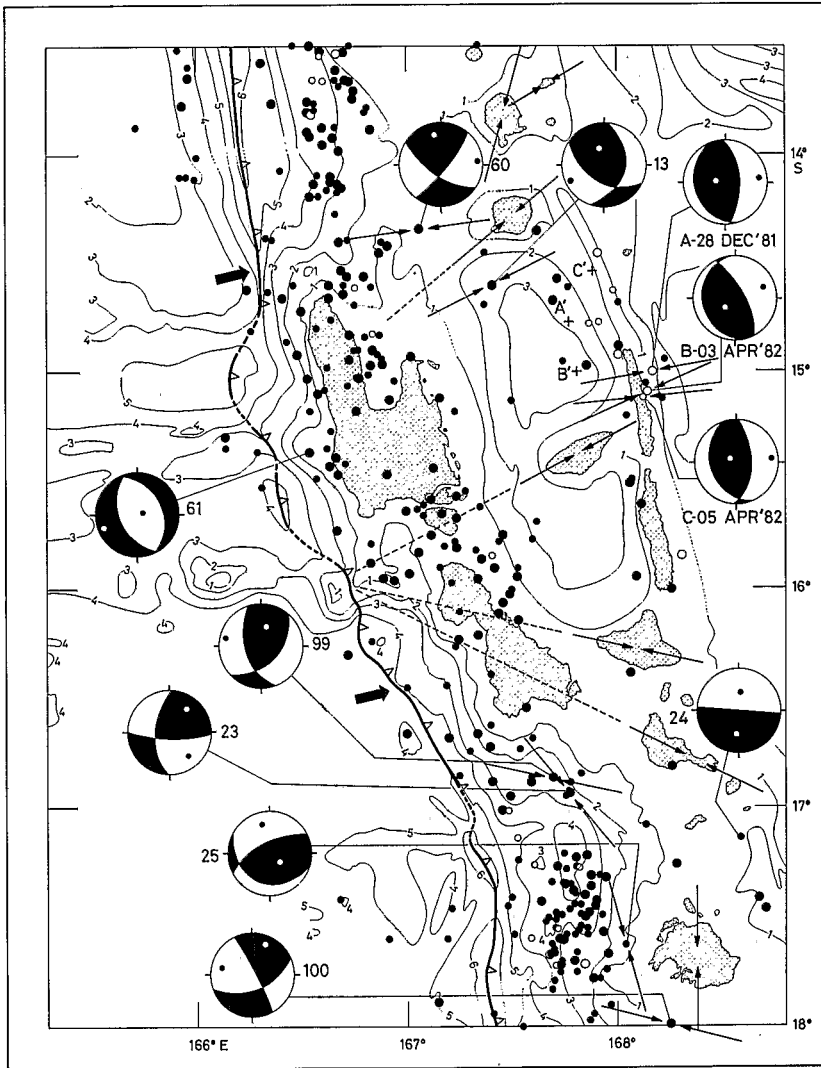
Fig. 11. Shallow (h < 80 km) moderate-sized earthquake distribution in the central New Hebrides. Names of the islands are indicated in Fig. 1. Filled circles indicate ISC data (01/1961 to 06/1979); open circles indicate monthly preliminary determination of epicenters (07/1979 to 05/1982) (the size of the circle is a function of the number (N) of stations used to locate the earthquake: large circle N > 100, small circle N > 50). Focal mechanism solutions of intraplate earthquakes are shown in lower hemisphere equal area projections with quadrants of compressional first motions blackened, compressional axes as filled circles and tensional axes as open circles: 23 from Pascal et al. (1978); 24, 25; 13 from Johnson and Molnar (1972); 60, 61 from Isacks et al. (1981); 99, 100 from Chatelain et al. (in press); A, B, C from PDE bulletin (crosses indicate centroïd locations). Light arrows indicate horizontal projections of maximum compressive stress axes from both the earthquake mechanism solutions and structural evidences (Roca, 1978; Katz and Daniel, 1981; Geological Survey of the New Hebrides, 1974). Large arrows indicate the direction of plate convergence. The broad line is the outer limit of the plate contact zone.

northward than southward. This local stress field appears, nevertheless, to be superimposed on a N76° normal stress field related to the subduction process (Roca, 1978).

Theoretically, a strong horizontal interaction generates compressive features or even strike–slip movements, such as those observed in a plane collision pattern. However, onshore geological studies of the highly deformed western and eastern
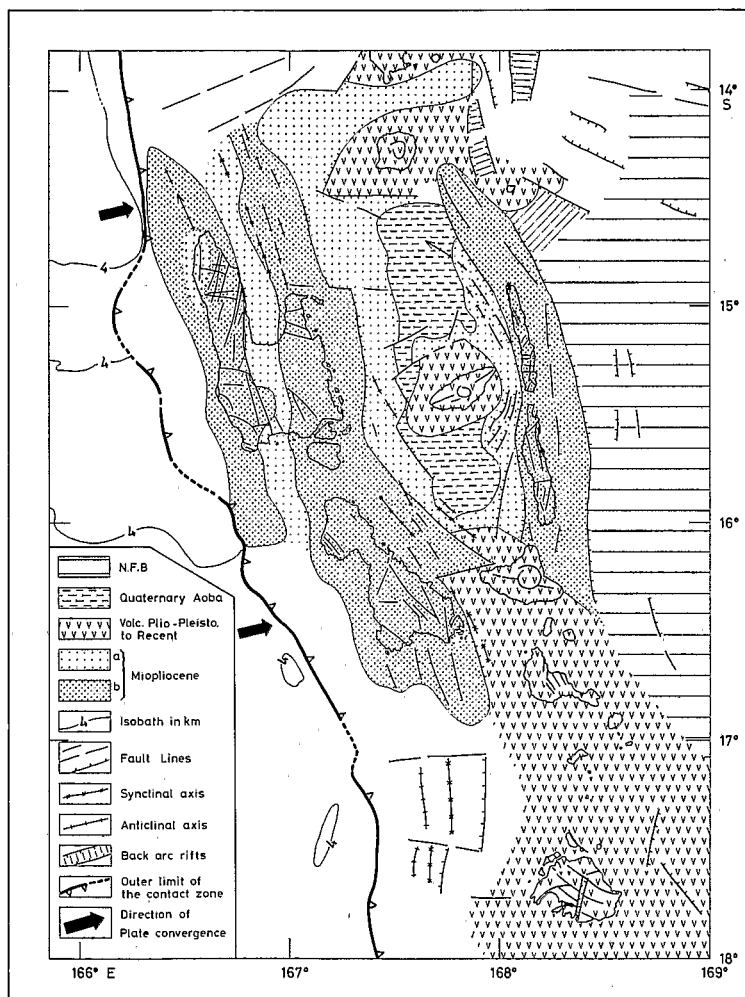

Fig. 12. Structural map after Katz and Daniel (1981); NFB = North Fiji Basin; Quaternary Aoba = Quaternary sedimentary cover, in South and North Aoba basins; Volc. Plio-Pleisto. to Recent = volcanic complexes Plio-Pleistocene to Recent related to the active central chain; Miopliocene = Mio-Pliocene sedimentary basin (a) only moderately deformed, (b) strongly deformed (Mio-Pliocene sediments are partially covered with quaternary sediments or reefs). Fault lines, synclinal axes and anticlinal axes on Santo, Malekula, Pentecost and Maewo after Carney and Macfarlane (1982).

belts (Fig. 12) by Robinson (1969), Mitchell (1966), Mallick and Neef (1974), Mallick and Greenbaum (1977), Carney (in press), as well as offshore geological results by Luyendyk et al. (1974), Ravenne et al. (1977), Katz and Daniel (1981), have not revealed any clear evidence of Plio-Pleistocene strike–slip movements, although many Miocene transcurrent faults could have been more recently re-activated, for example those in the west Santo area. Some arc transverse crustal fractures representing major structural and geological discontinuities have been reported by Karig and Mammerickx (1972), Mallick (1973), Luyendyk et al. (1974), Dugas et al. (1977), and Carney and Macfarlane (1982). These transverse directions are emphasised by strong magnetic trends deduced both from Malahoff's (1970) results and from Fig. 4 for the North Aoba basin border. This latter area has been recently reactivated by movements with a strike–slip component (Isacks et al., 1981).

Geological studies show some evidence for recent E–W compressive deformation. In the central basin two units can be defined. A recent and relatively undisturbed sedimentary layer overlies a thick Miocene–Pliocene series with angular unconform-ity to the west and more or less conformable to the east. The lower unit extending from the western chain to the top of the eastern chain is down-flexed (fig. 2 in Carney and Macfarlane, 1980). In the eastern part of the basin, a few low-amplitude folds affect the whole sedimentary section (Fig. 12). In the eastern chain the sedimentary section appears to have suffered severe faulting and folding. In our opinion, this makes it difficult to separate the different units and tectonic episodes. These observations suggest that the eastern chain has been deformed, uplifted, and tilted to the west under horizontal and vertical stresses. Moreover, some of this deformation must be Recent, since Recent sediments are deformed. Low-amplitude folds affecting Quaternary limestone terraces were reported on Pentecost (Mallick and Neef, 1974) and Maewo (Carney, in press). They may overlap areas of maximum uplift in the basement (Mallick and Neef, 1974), but they also have resulted from recent lateral compression. Carney and Macfarlane (1982) discussed the polyphase tectonic development of these islands. They suggested that faulting, flexuring, and westward tilting observed onshore were developed during uplift from the post-3 Ma major tectonic episode. As for the active tectonics of the western chain, these are mainly controlled by vertical movements, except for the inner wall deformation and some late reversed movements reported in western Santo. Thus, although a radiating horizontal compressive stress field can be established in the central arc, related horizontal deformation is not widely distributed, except along the eastern belt. This last assertion is emphasized by the overall configuration of the eastern chain, which is that of an arcuate belt, convex to the east (Fig. 12).

On the other hand, vertical deformation is widely developed throughout much of the New Hebrides, which has a mean Quaternary uplift rate of 1 mm/yr. The highest Holocene uplift rates have been recorded in the western and eastern belts (Mallick and Neef, 1974; Jouannic et al., 1980). In fact, the western chain Plio-Quaternary elevation began 2 Ma ago (Carney and Macfarlane, 1982) and mainly

predates the eastern belt elevation (Luyendyk et al., 1974), which began in the early Pleistocene (Mallick and Neef, 1974). Both chains emerged in the late Pleistocene, and reached about 1800 m on Santo (Carney and Macfarlane, 1980) and 950 m on Pentecost. The results of this tectonic episode are also apparent on seismic reflection profiles over both sides of the Aoba Basin. Thus, although the basement of the basin is of early Miocene age, the present basin-like structure results from the Pliocene–Pleistocene tectonic episode (Ravenne et al., 1977). Since the d'Entrecasteaux Zone has been subducting and generating additional horizontal compressive stresses for at least 2 Ma (Pascal et al., 1978), the Recent tectonic history of the central arc can be related temporally to the d'Entrecasteaux Zone–arc interaction. A clear correlation has been pointed out by Taylor et al. (1980) on the plate edge. In fact, the Santo and Malekula Quaternary coral terraces are segmented into blocks separated by inferred tilt discontinuities that trend perpendicular to the arc. These are located in front of both of the subducting d'Entrecasteaux ridges.

(2) *Seismicity.* Recent studies by Pascal et al. (1978), and Isacks et al. (1981) have shown that, on a regional scale, the overall configuration of the Benioff Zone beneath the New Hebrides is relatively uniform, although some unusual features of the intermediate and shallow seismicities were noted in the central arc and mainly related to the subduction of the d'Entrecasteaux Zone (Choudhury et al., 1975; Chung and Kanamori, 1978a; Isacks et al., 1981; Louat et al., 1982; Marthelot, 1983). In this section, we shall consider the tectonic implications of the shallow seismicity of the central arc, referring mainly to the results of Isacks et al. (1981).

The shallow seismicity is basically concentrated along the edge of the upper plate, as is normally expected in a subduction zone (Fig. 11). A feature of the thrust zone between latitude 14° and 16°S is the large scattering of the slip vector azimuths with respect to the mean azimuth of N76°. Moreover, this seismic activity has been shown to occur within segments delimiting rupture zones, whose boundaries coincide with the tilt discontinuities of the uplifted Quaternary terraces. This may be correlated with the subduction of the North d'Entrecasteaux Ridge and the South d'Entrecasteaux Chain. In addition, although Kelleher and McCann (1976) concluded that large shallow earthquakes occurred less frequently where ridges are consumed than along other segments of subduction, Louat et al. (1982) pointed out that the largest sequences of shallow earthquakes of the New Hebrides occurred in the Santo region, which is evidence for a strong interplate coupling, owing to the peculiar structure of the frontal arc. In the Aoba Basin, the seismic activity is low, but it significantly increases in the adjacent back-arc region (Fig. 11), which emphasises the active tectonics of the eastern chain and coincides with the transition zone between an old and presumably thick island arc crust and the recent and thin North Fiji Basin oceanic crust.

Reliable focal mechanism solutions from Johnson and Molnar (1972), Pascal et al. (1978), Isacks et al. (1981), Chatelain et al. (in press) and from the monthly bulletins of the PDE (Preliminary Determination of Epicentres), presumed to be for

earthquakes located within the upper plate are reported in Fig. 11.

The pattern obtained from horizontal projection of compressive stress axes is consistent with the stress field deduced from the geological data. The focal mechanism analysis indicates mainly three pure types of faulting: normal faulting (events 61–24), thrust faulting (events A–B–C), strike-slip faulting (events 60–23–100). Three others are of intermediate type with a strike-slip component (events 13–25–99). In plan view, the distribution of the strike-slip and thrust types looks organised. In fact, in the light of the above mentioned crustal fractures that traverse the arc, the fault-plane solutions can be interpreted as transverse sinistral strike-slip along the northern boundary of the Aoba Basin, and transverse dextral strike-slip in the south Malekula area (Fig. 13). This suggests a relative displacement eastward of the central block bounded by these features, thus accounting for the compressive thrust zone along its eastern border, which is considered as a weak transition zone.

(3) *Discussion.* This strain pattern across the arc (Fig. 13) is analogous to the theoretical strain field (Fig. 14) resulting from the indentation of a long and narrow rigid-plastic body (i.e. the arc) by a flat rigid die (i.e. the d'Entrecasteaux Zone and
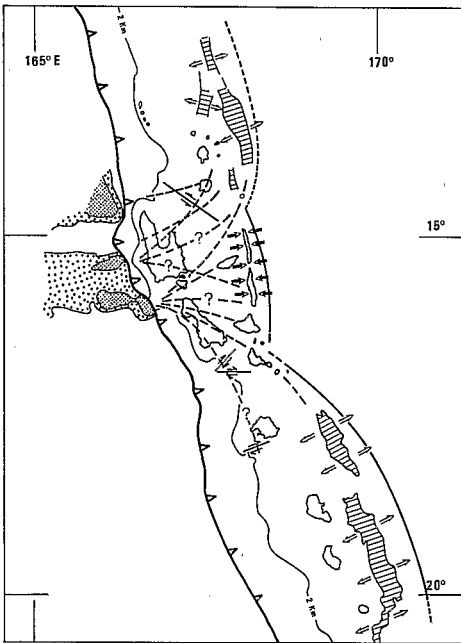

Fig. 13. Sketch map of the stress strain pattern along the arc inferred from the model developed in this paper. Dotted areas are DEZ and WTM. Ruled areas are rear-arc troughs. Solid lines show western and eastern boundaries of the arc. Long dashed lines indicate the radiating stress pattern. Strike–slip faults from earthquake mechanism solutions are represented with sense of shear. Filled arrows show the zone of compressive stress and open arrows show tensional zone. 2 km isobath emphasises the protruded shape of the Central block.

perhaps the southeastern corner of the West Torres Massif). The pattern of such deformation depends upon the boundary conditions. If the indented body is of finite width, the plastic region becomes wider as the ratio of the width of the body to the width of the indenting die increases (Tapponnier and Molnar, 1976). If this ratio is small, displacements of the indenting die, producing a motion of the adjacent triangular block, tend to cut the narrow body in two parts. Applied to the New Hebrides case, this process would force the north and the south section of the arc to separate rigidly away from the central block. This should develop strike–slip motion along the lateral boundaries of the central block and along the eastern boundary of the north and south sections of the New Hebrides arc. Moreover, this theory predicts thrust faulting, crustal thickening and uplifting in front of the indenting die, and also secondary tension, associated normal faulting, and crustal extension in the lateral regions under certain boundary conditions (Molnar and Tapponnier, 1977).

Although, on the scale of d'Entrecasteaux Zone collision, physical indentation does not really occur, the coupling between lower and upper plates in the Santo–Malekula area appears to be strong enough to generate horizontal compressive stresses that are transmitted across the central block towards its boundaries. The main body of the central block seems to behave as a "dead structure", and geological evidence suggests that it is poorly deformed by horizontal stresses, whereas its edges suffer much more deformation.

In front of d'Entrecasteaux Zone the accumulated energy is mainly released in the form of thrusts and uplift—firstly at the present subduction zone, and secondly
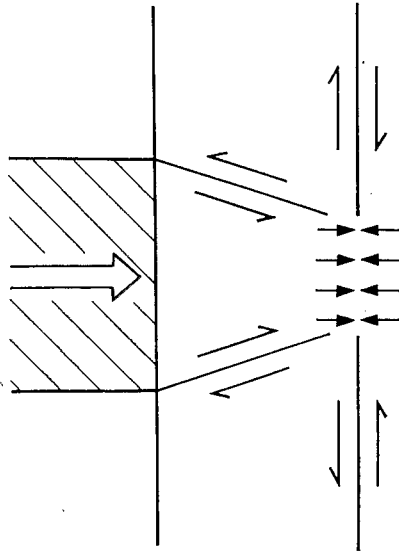

Fig. 14. Diagrammatic strain pattern in the particular case of indentation of a finite width body by a flat rigid die (after Molnar and Tapponnier, 1975).

along the weak eastern belt area. In both lateral areas, energy is partly released in the form of strike-slip deformation. This process is facilitated by pre-existing fault zones of the right orientation in the North Aoba area, and also by the narrowness of the arc in the South Malekula–Epi area. In fact, transverse crustal fractures may not affect the whole width of the arc.

The thrust focal-mechanism solutions support the assertion of Isacks et al. (1981) that the eastern chain is now under horizontal compression. Considering the early Miocene age of the Pentecost–Maewo basement, the style of the Recent deformation affecting the west flank of the ridge, and the existence of the young adjacent North Fiji basin, we interpret the A–B–C events (Fig. 11) as evidence of thrusting along a west-dipping moderate-angle fault plane, thus accounting for the recent and active uplift of the eastern chain. This is consistent west-dipping subduction.

Chung and Kanamori (1978b) proposed a mechanical model that satisfactorily explains the general uplift of the western chain as well as the associated depression of the Aoba Basin, by simply applying, at the edge of a semi-infinite elastic plate, a lone upward loading due to the buoyancy of the d'Entrecasteaux Zone. However, it appears that horizontal loading would better fit the observations. We suggest that both the western and the eastern chain uplift could thus be explained by applying an upward loading at each of the two opposite edges of an elastic plate of finite width. The right balance between the loading should be calculated to account for the positive edge deformations limiting a down bending zone, the Aoba Basin. The upward loading needed at the eastern edge could be explained by the initial
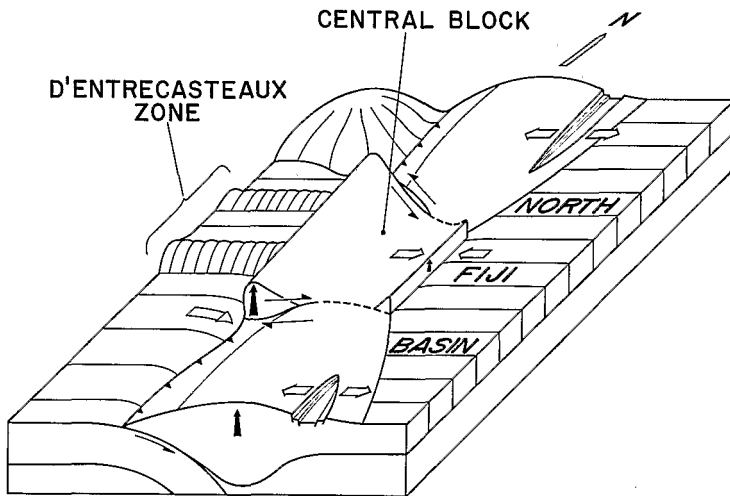

Fig. 15. Three-dimensional schematic model, illustrating the main tectonic features of the New Hebrides resulting from the subduction/collision process of the d'Entrecasteaux Zone. The figure is not to scale. The frontal cross section is based on seismic refraction (Pontoise et al., 1982) and gravity data (Collot and Malahoff, 1982).

conditions of the thrusting process. Thus, the central New Hebrides block appears to be elastically bent, mainly under vertical forces, and, at the same time, plastically deformed and pushed eastward as a wedge under horizontal compression (Fig. 15).

Although there is no evidence for strike-slip motion along the eastern border of the north and south parts of the arc, crustal extension is well developed there (Dubois et al., 1975; Coudert et al., 1981). It can be seen from Fig. 13, that the radiating compressive stresses developed in the central arc become parallel to the arc trend in the northern and southern lateral sections of the arc. This agrees with the main direction of extension developed in these rear-arc regions of normal faulting. Thus, considering the border between the arc and the North Fiji Basin as a weak zone, we suggest that the observed plane-strain field could be mainly the result of an elastic-plastic mechanical behaviour of the arc when deformed under horizontal loading. Horizontal elastic strain should have affected the arc. This strain may be compared to that resulting from the downward loading of an elastic beam overlying
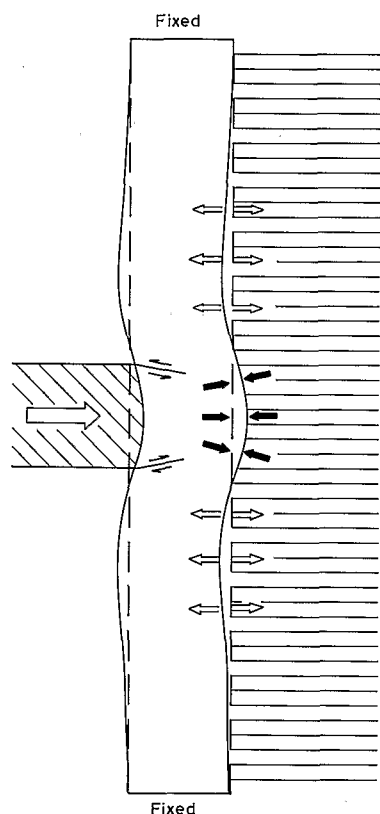

Fig. 16. Diagrammatic plane strain pattern for an elastic-plastic beam when deformed by a horizontal loading. Filled arrows: compressive zone; open arrows: tensional zones.

a fluid substratum. Such a theoretical problem has been widely discussed by Walcott (1970). In fact, to apply this model in plan view to the New Hebrides case, we must assume that North Fiji Basin, an active marginal sea, is weaker than the arc. This process should generate lateral secondary tension as indicated in Fig. 16. Thus, according to this hypothesis, the rear-arc troughs recognised in both northern and southern sections of the arc could be a direct consequence of the subduction/collision of the d'Entrecasteaux Zone. Even here, when the yield stress is reached, plastic deformation occurs.

In the central arc the minimum compressive stress ($\sigma 3$) is vertical when the maximum compressive stress ($\sigma 1$) is cross horizontal. At the rear of both north and south sections of the arc $\sigma 3$ deduced from the tensional tectonics is cross horizontal and $\sigma 2$ is the maximum compressive horizontal stress.

CONCLUSION

The compressive style of regional tectonics in the central New Hebrides is mainly related to the collision and underthrusting of a complex ridge zone with an old protruding island arc block: an exceptional situation, rarely encountered in intra-oceanic subduction zones. Resulting deformations are of two orders: local deformation due to the indentation of the upper plate edge by topographic features of the downgoing plate and large scale deformation throughout the central New Hebrides, due to the local and strong coupling between the plates.

In both north and south sections of the arc, tensional tectonic styles have been reported. The two drastically different stress regimes allow the New Hebrides rear-arc region to be divided into three main sections: the northern New Hebrides—a tensional zone from the north to 14°30'; the central New Hebrides—a compressive zone between 14°30' and 17°; the southern New Hebrides—a tensional zone between 17° and 20°30'. Besides, the rear-arc region north of Hunter and Matthew islands in the southernmost New Hebrides is interpreted by Maillet et al. (in prep.) as a complex zone of sinistral shear.

As discussed above, the collision process at the plate boundary can be invoked to account for the compressive and lateral tensional stress regimes. However, it should be noted that the incompletely known spreading pattern of the associated North Fiji Basin might also have a significant influence on both the active tectonics of the New Hebrides and the general stress block distribution.

REFERENCES

Andrews, J.E., Packham, G.H. et al., 1975. Site 286. In: Initial Reports of the Deep Sea Drilling Project, Vol. 30. US Government Printing Office, Washington, D.C., pp. 69–131.

Carney, J., in press. Geology of Maewo. Reg. Rep. Geol. Surv. New Hebrides.

Carney, J. and Macfarlane, A., 1980. A sedimentary basin in the Central New Hebrides. UN ESCAP–C-COP/SOPAC. Tech. Bull. 3, pp. 109–120.

Carney, J. and Macfarlane, A., 1982. Geological evidence bearing on the miocene to recent structural evolution of the New Hebrides arc. Tectonophysics, 87: 147–175.

Chatelain, J.L., Cardwell, R.K. and Isacks, B.L., 1983. Expansion of the aftershock zone following the Vanuatu (New Hebrides) earthquake of 15 July 1981. Geophys. Res. Lett., 10 (5): 385–388.

Chatelain, J.L., Cardwell, R.K., Isacks, B.L. and Bevis, M., in press. Detailed study of seismicity patterns and tilt associated with a sequence of moderate size interplate earthquakes in the New Hebrides Island Arc. J. Geophys. Res.

Choudhury, M.A., Poupinet, G. and Perrier, G., 1975. Shear velocity from differential travel times of short-period ScS–P in New Hebrides, Fiji–Tonga and Banda sea regions. Bull. Seismol. Soc. Am., 65: 1787–1796.

Chung, W.Y. and Kanamori, H., 1978a. Subduction process of a fracture zone and aseismic ridges—the focal mechanism and source characteristics of the New Hebrides earthquake of 1969 January 19 and some related events. Geophys. J.R. Astron. Soc., 54: 221–240.

Chung, W.Y. and Kanamori, H., 1978b. A mechanical model for plate deformation associated with aseismic ridge subduction in the New Hebrides arc. Tectonophysics, 50 (1): 29–40.

Colley, H. and Warden, A.J., 1974. Petrology of the New Hebrides. Geol. Soc. Am. Bull., 85: 1635–1646.

Collot, J.Y. and Malahoff, A., 1982. Anomalies gravimétriques et structure de la zone de subduction des Nouvelles-Hébrides. In: Equipe de Géologie–Géophysique du Centre ORSTOM de Nouméa. Contribution à l'étude géodynamique du Sud-Ouest Pacifique. Trav. Doc. ORSTOM, No. 147, pp. 91–110.

Collot, J.Y., Missegue, F. and Malahoff, A., 1982. Anomalies gravimétriques et structure de la croûte dans la région de la Nouvelle-Calédonie: enracinement des péridotites. In: Equipe de Géologie-Géophysique du Centre ORSTOM de Nouméa. Contribution à l'étude géodynamique du Sud-Ouest Pacifique. Trav. Doc. ORSTOM, No. 147, pp. 499–564.

Coudert, E., Isacks, B.L., Baranzangi, M., Louat, R., Cardwell, R., Chen, A., Dubois, J., Latham, G. and Pontoise, B., 1981. Spatial distribution and mechanisms of earthquakes in the Southern New Hebrides arc from a temporary land and ocean bottom seismic network and from world-wide observations. J. Geophys. Res., 86: 5905–5925.

Daniel, J. and Katz, H.R., 1981. D'Entrecasteaux Zone, trench and western chain of the central New Hebrides island arc: their significance and tectonic relationship. Geo-Mar. Lett., 1: 213–219.

Daniel, J., Jouannic, C., Larue, B. and Recy, J., 1977. Interpretation of d'Entrecasteaux Zone (north of New Caledonia). Int. Symp. on Geodynamics in South-West Pacific, Noumea 1976. Technip, Paris, pp. 117–124.

Dickinson, W.R. and Seely, D.R., 1979. Structure and stratigraphy of forearc regions. Am. Assoc. Petrol. Geol. Bull., 63 (1): 2–31.

Dubois, J., Dugas, F., Lapouille, A. and Louat, R., 1975. Fossés d'effondrements en arrière de l'arc des Nouvelles-Hébrides. Mécanismes proposés. Rev. Geogr. Phys. Géol. Dyn., 17(1): 73–94.

Dugas, F., Dubois, J., Lapouille, A., Louat, R. and Ravenne, C., 1977. Structural characteristics and tectonics of an active island arc: The New Hebrides. Int. Symp. on Geodynamics in South-West Pacific, Noumea 1976. Technip, Paris, pp. 79–90.

Dupont, J., 1979. Le système d'arc insulaire des Tonga et Kermadec: deux morphologies différentes, une seule zone de subduction (Pacifique Sud). C.R. Acad. Sci. Paric, Sér. D, 289: 245–248.

Dupont, J., 1982. Morphologie et structures superficielles de l'arc insulaire des Tonga–Kermadec. In: Equipe de Géologie—Géophysique du Centre ORSTOM de Nouméa. Contribution à l'étude géodynamique du Sud-Ouest Pacifique. Trav. Doc. ORSTOM, No. 147, pp. 263–282.

Dupont, J. and Recy, J., 1981. Le Sud-Ouest Pacifique: données structurales. In: Atlas de la Nouvelle-Calédonie et Dépendances. ORSTOM, Paris.

Faleyev, V.I., Udintsev, G.B., Agapova, G.V., Domanitskiy, Ye.A. and Marova, N.A., 1977. Data on the maximum depths of trenches in the world ocean. Oceanology, 17 (3): 311–313.

Geological Survey of the New Hebrides, 1974. Geological map of the New Hebrides Condominium.

Gorton, M.P., 1974. The geochemistry and geochronology of the New Hebrides. Ph.D. Thesis, Australian National University, Canberra, A.C.T., 300 pp. (unpublished).

Gorton, M.P., 1977. The geochemistry and origin of Quaternary volcanism in the New Hebrides. Geochim. Cosmochim. Acta, 41: 1257–1270.

Ibrahim, A.K., Pontoise, B., Latham, G.V., Larue, B.M., Chen, T., Isacks, B.L., Recy, J. and Louat, R., 1980. Structure of the New Hebrides arc-trench system. J. Geophys. Res., 85: 253–266.

Isacks, B.L., Cardwell, R.K., Chatelain, J.L., Barazangi, M., Marthelot, J.M., Chinn, D. and Coudert, E., 1981. Seismicity and tectonics of the central New Hebrides island arc. In: Earthquake Prediction—An International Review. Maurice Ewing Ser., Am. Geophys. Union, 4.

Johnson, T. and Molnar, P., 1972. Focal mechanisms and plate tectonics of the Southwest Pacific. J. Geophys. Res., 77: 5000–5032.

Jouannic, C., 1975. Carte bathymétrique de la région Rennell–d'Entrecasteaux. ORSTOM, Nouméa (unpublished).

Jouannic, C., Taylor, F.W., Bloom, A.L. and Bernat, M., 1980. Late Quaternary uplift history from emerged reef terraces on Santo and Malekula islands, central New Herbides island arc. UN ESCAP-CCOP/SOPAC Tech. Bull., 3: 91–108.

Jouannic, C., Taylor, F.W. and Bloom, A.L., 1982. Sur la surrection et la déformation d'un arc jeune: l'arc des Nouvelles Hébrides. In: Equipe de Géologie–Géophysique du Centre ORSTOM de Nouméa. Contribution à l'étude géodynamique du Sud-Ouest Pacifique. Trav. Doc. ORSTOM, No. 147, pp. 223–246.

Karig, D.E. and Mammerickx, J., 1972. Tectonic framework of the New Hebrides island arc. Mar. Geol., 12: 187–205.

Katz, H.R., 1981. Report on interpretation of seismic profiling data collected on the VAUBAN cruise in Vanuatu waters. Tech. Sect. CCOP/SOPAC, Suva, Fiji (unpublished).

Katz, H.R., 1982. Southwest Pacific island arcs: sedimentary basins and petroleum prospects in the New Hebrides and Solomons. Trans. Vol.; Third Circum-Pacific Energy and Mineral Resources Conf., August 22–28, 1982, Honolulu.

Katz, H.R. and Daniel, J., 1981. Structural map of the New Hebrides island arc (unpublished).

Kelleher, J. and McCann, W., 1976. Buoyant zones, great earthquakes, and unstable boundaries of subduction. J. Geophys. Res., 81: 4885–4896.

Lapouille, A., 1982. Etude des bassins marginaux fossiles du Sud-Ouest Pacifique: bassin Nord d'Entrecasteaux, bassin Nord-Loyauté, bassin Sud-Fidjien. In: Equipe de Géologie–Géophysique du Centre ORSTOM de Nouméa. Contribution à l'étude géodynamique du Sud-Ouest Pacifique. Trav. Doc. ORSTOM, No. 147, pp. 409–438.

Lapouille, A. and Dugas, F., 1975. Geological evolution of New Hebrides and loyalty areas. Bull. Aust. Soc. Explor. Geophys., 6 (2/3): 52.

Louat, R., Daniel, J. and Isacks, B.L., 1982. Sismicité de l'arc des Nouvelles Hébrides. In: Equipe de Géologie–Géophysique du Centre ORSTOM de Nouméa. Contribution à l'étude géodynamique du Sud-Ouest Pacifique. Trav. Doc. ORSTOM, No. 147, pp. 111–148.

Luyendyk, B.P., Bryan, W.B. and Jezek, P.A., 1974. Shallow structure of the New Hebrides island arc. Geol. Soc. Am. Bull., 85: 1287–1300.

Maillet, P., Monzier, M., Selo, M. and Storzer, D., 1983. The d'Entrecasteaux Zone (South-West Pacific) —a petrological and geochronological reappraisal. Mar. Geol., 53: 179–198.

Maillet, P., Monzier, M. and Lefevre, C., in prep. Petrology of Matthew and Hunter volcanoes, south New Hebrides island arc.

Malahoff, A., 1970. Gravity and magnetic studies of the New Hebrides island arc. New Hebrides Condominium Geological Survey, Port-Vila, 67 pp.

Mallick, D.I.J., 1973. Some petrological and structural variations in the New Hebrides. In: P.J. Coleman (Editor), The Western Pacific: Island Arcs, Marginal Seas, Geochemistry. University of Western Australia Press, Nedlands, W.A., pp. 193–211.

Mallick, D.I.J., 1975. Development of the New Hebrides archipelago. Philos. Trans. R. Soc. London, Ser. B, 272: 277–285.

Mallick, D.I.J. and Greenbaum, D., 1977. Geology of southern Santo. N.H. Cond. Surv. Reg. Rep., 84 pp.

Mallick, D.I.J. and Neef, G., 1974. Geology of Pentecost. Reg. Rep. Geol. Surv. New Hebrides, 103 pp.

Mammerickx, J., Chase, T.E., Smith, S.M. and Taylor, I.L., 1971 (revised 1974). Bathymetry of the South Pacific. Chart No. 12. Scripps Institution of Oceanography, Woods Hole, Mass.

Marthelot, J.M., 1983. Seismicity and attenuation in the central Vanuatu (New Hebrides arc): a new interpretation of the effect of subducting the d'Entrecasteaux fracture zone. Ph.D. Thesis, Cornell Univ., Ithaca, N.Y.

Mitchell, A.H.G., 1966. Geology of South Malekula. N.H. Cond. Geol. Surv. Rep. No. 3, 42 pp.

Mitchell, A.H.G. and Warden, A.J., 1971. Geological evolution of the New Hebrides island arc. J. Geol. Soc. London, 127: 501–529.

Molnar, P. and Tapponnier, P., 1975. Cenozoic tectonics of Asia: effects of a continental collision. Science, 189 (4201): 419–426.

Molnar, P. and Tapponnier, P., 1977. Relations of the tectonics of eastern China to the Indian-Eurasia collision: application of slip-line field theory to large-scale continental tectonics. Geology, 5: 212–216.

Moretti, I., 1983. Subduction des rides aseismiques. Thèse 3e cycle, Université de Paris-Sud, Paris.

Nakamura, K. and Uyeda, S., 1980. Stress gradient in arc-back arc regions and plate subduction. J. Geophys. Res., 85: 6419–6428.

Pascal, G., 1974. Contribution à l'étude de la sismicité des Nouvelles-Hébrides. Thèse Doctorat Sciences, Faculté des Sciences de l'Université de Paris VI, Paris, 133 pp.

Pascal, G., Isacks, B.L., Barazangi, M. and Dubois, J., 1978. Precise relocations of earthquakes and seismotectonics of the New Hebrides island arc. J. Geophys. Res., 83: 4957–4973.

Pontoise, B., Latham, G.V., Daniel, J., Dupont, J. and Ibrahim, A.K., 1980. Seismic refraction studies in the New Hebrides and Tonga area. UN ESCAP-CCOP/SOPAC Tech. Bull. 3, pp. 47–58.

Pontoise, B., Latham, G.V. and Ibrahim, A.K., 1982. Sismique réfraction: structure de la croûte aux Nouvelles-Hébrides. In: Equipe de Géologie–Géophysique du Centre ORSTOM de Nouméa. Contribution à l'étude géodynamique du Sud-Ouest Pacifique. Trav. Doc. ORSTOM, No. 147, pp. 79–90.

Ravenne, C., Pascal, G., Dubois, J., Dugas, F. and Montadert, L., 1977. Model of a young intra-oceanic arc: the New Hebrides island arc. International Symposium of Geodynamics in South-West Pacific, Nouméa 1976. Technip, Paris, pp. 63–78.

Robinson, G.P., 1969. The geology of North Santo. N.H. Geol. Surv. Reg. Rep., 77 pp.

Roca, J.L., 1978. Contribution à l'étude pétrologique et structurale des Nouvelles-Hébrides. Thèse de 3e cycle, Université des Sciences et Techniques du Languedoc, Montpellier, 158 pp.

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Tapponnier, P. and Molnar, P., 1976. Slip-line field theory and large-scale continental tectonics. Nature, 264: 319–324.

Taylor, F.W., Isacks, B.L., Jouannic, C., Bloom, A.L. and Dubois, J., 1980. Coseismic and Quaternary vertical tectonic movements, Santo and Malekula islands, New Hebrides island arc. J. Geophys. Res., 85: 5367–5381.

Vogt, P.R., Lowrie, A., Bracey, D.R. and Hey, R.N., 1976. Subduction of aseismic oceanic ridges: effects on shape, seismicity, and other characteristics of consuming plate boundaries. Geol. Soc. Am., Spec. Pap., 172: 59 pp.

Walcott, R.I., 1970. Flexure of the lithosphere at Hawaii. Tectonophysics, 9: 435–446.

Watts, A.B., Weissel, J.K. and Davey, F.J., 1977. Tectonic evolution of the South Fiji marginal basin. In: M. Talwani and W.C. Pitman III (Editors), Island Arcs, Deep Sea Trenches and Back-Arc Basins. Maurice Ewing Ser., Am. Geophys. Union, 2: 419–427.

Weissel, J.K., Watts, A.B., Lapouille, A., Karner, G. and Jongsma, D., 1977. Preliminary results from recent geophysical investigations in marginal basins of Melanesia. EOS, Trans. Am. Geophys. Union, 58: 504.

Weissel, J.K., Watts, A.B. and Lapouille, A., 1982. Evidence for late Paleocene to late Eocene seafloor in the Southern New Hebrides basin. Tectonophysics, 87: 243–251.