Collot, J.-Y., Greene, H. G., Stokking, L. B., et al., 1992
Proceedings of the Ocean Drilling Program, Initial Reports, Vol. 134

1. INTRODUCTION[1]

Shipboard Scientific Party[2]

OVERVIEW

The New Hebrides Island Arc is the product of intra-oceanic subduction, possible reversal of subduction polarity, and ridge-arc collision, all of which has taken place during the Cenozoic. Leg 134 focused on three important and poorly understood aspects of this arc system, namely:

1. The style of deformation occurring in the region of the collision between the d'Entrecasteaux Zone (DEZ) and the New Hebrides Island Arc (Fig. 1). More specifically we compare the features of deformation in two different types of ridges in the DEZ that are impinging upon the arc: the continuous North d'Entrecasteaux Ridge (NDR) and the seamounts of the South d'Entrecasteaux Chain (SDC) (Fig. 2). Impingement of the DEZ against the arc has greatly altered the arc's morphology and structure. To determine the recycling of lithosphere, or the transfer of material from one plate to another, a series of holes (Fig. 2) were drilled through the sediments, carbonates, and volcanic rocks of the forearc (Sites 827, 829, and 830), of the NDR on the Pacific plate (Site 828), and of the Bougainville Guyot of the SDC (Site 831).
2. The evolution of the magmatic arc in relation to a possible reversal of subduction during the Neogene. To investigate the evolution of intra-arc basins and to determine whether or not subduction polarity reversed, two sites were drilled in one of two summit basins of the arc, the intra-arc North Aoba Basin (Sites 832 and 833; Fig. 2). A major discordance in the basin fill may be contemporaneous with the beginning of collision of the DEZ with the arc, providing one of the best estimates of the age of subduction polarity reversal and initiation of collision. The holes drilled in the North Aoba Basin show the provenance, age, paleobathymetry, and lithology of basin fill, from which the rate and timing of basin subsidence can be derived.
3. Dewatering of the forearc and subducted lithosphere, to be investigated indirectly from the composition of the forearc crust and directly from the analyses of fluids and chemical precipitates from the forearc.

Most of the background information in the following paragraphs is after Greene and Wong (1988). The time scale used is that of Berggren et al. (1985).

THE NEW HEBRIDES ISLAND ARC

Tectonic Setting

The New Hebrides Island Arc system extends for a distance of 1700 km from the Santa Cruz Islands in the north to Matthew and Hunter islands in the south (Fig. 1). This arc lies in the middle of a complex system of active volcanic arcs that extends from Papua New Guinea southward through the Solomon Islands to Vanuatu, where the New Hebrides Island Arc is displaced eastward along the Hunter Fracture Zone to Tonga (Fig. 1). From Tonga the arc system continues southward through the Kermadec Islands to New Zealand. Along most of the arc system, the Pacific plate is being subducted under the Australia-India plate. From Papua New Guinea to Vanuatu, however, the arc system defines a major plate boundary where the Australia-India plate is being subducted beneath the Pacific plate and the North Fiji Basin.

The summit platform in the central part of the New Hebrides Island Arc, from the Torres Islands southward, has an average width of 200 km and is approximately 1450 km long. It is bounded on the west by the New Hebrides Trench except in the central New Hebrides Island Arc region where no physiographic expression of subduction is seen (Fig. 3). The trench correlates with a fairly continuous east-dipping Benioff zone (Karig and Mammerickx, 1972; Pascal et al., 1978; Louat et al., 1988; Macfarlane et al., 1988). The physiographic trench is clogged by the DEZ, which is composed of ridges and basins (Daniel, 1978; Collot et al., 1985; Fisher, 1986; Fisher et al., 1986). The oblique collision of the DEZ with the central New Hebrides Island Arc (Pascal et al., 1978; Isacks et al., 1981) has produced a complex structural pattern. Differential uplift of the arc platform has further complicated the structure through the formation of horsts and grabens, which influenced the sedimentary depositional history of the central New Hebrides Island Arc (Carney and Macfarlane, 1982; Johnson and Greene, 1988; Greene and Johnson, 1988).

The central New Hebrides Island Arc has been divided into three distinct provinces of island-arc rocks, which are, from oldest to youngest, the Western Belt, the Eastern Belt, and the Central Chain (Mitchell and Warden, 1971; Mallick, 1973; Carney and Macfarlane, 1982; Carney et al., 1985; Macfarlane et al., 1988). These provinces apparently formed during three temporally and spatially overlapping volcanic episodes that occurred since the late Oligocene. The Western Belt is composed of the upper Oligocene to middle Miocene basement rocks of Malakula, Espiritu Santo, and the Torres islands. The Eastern Belt consists of the upper Miocene to lower Pliocene basement rocks of Pentecost and Maewo islands. The Central Chain comprises the Pleistocene volcanic islands of Ambrym, Aoba, and Santa Maria (Figs. 2 and 3).

Geologic History

Geologic and paleomagnetic evidence suggests that the initial development of the New Hebrides Island Arc took place behind the upper Eocene Vitiaz frontal arc. This tholeiitic frontal arc was then continuous with the arcs of Tonga and the Solomon Islands (outer Melanesian arc) above a west-dipping subduction zone (e.g., Chase, 1971; Karig and Mammerickx, 1972; Gill and Gorton, 1973; Carney and Macfarlane, 1978; Falvey, 1978; Kroenke, 1984; Falvey and Greene, 1988). A later relocation of the plate boundary to a position west of the Vitiaz arc and a possible reversal in subduction direction were accompanied by opening of the North Fiji Basin and

[1] Collot, J.-Y., Greene, H. G., Stokking, L. B., et al., 1992. *Proc. ODP, Init. Repts.,* 134: College Station, TX (Ocean Drilling Program).
[2] Shipboard Scientific Party is as given in the list of participants preceding the contents.

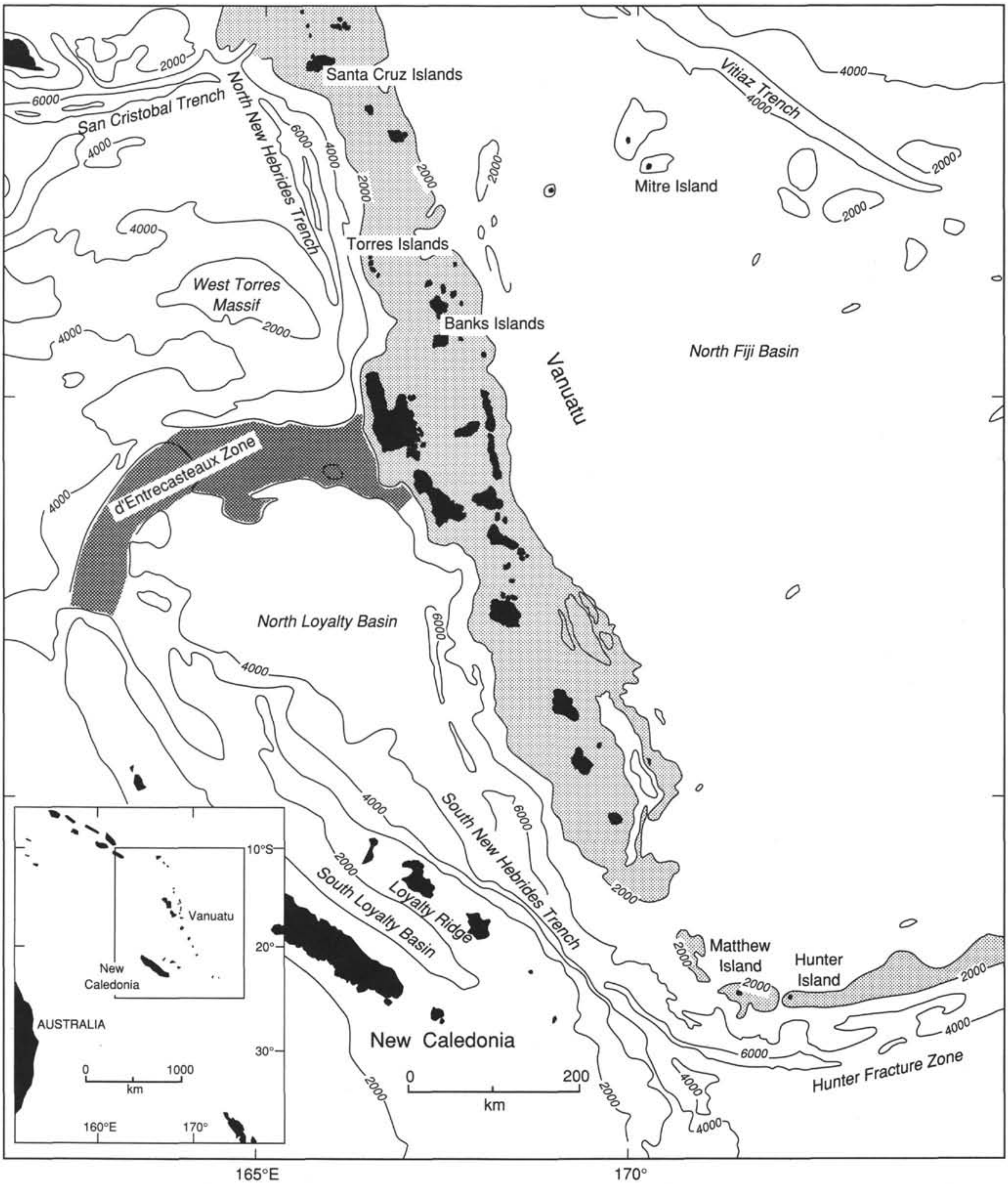

Figure 1. Regional bathymetry (in meters) of the southwest Pacific (after Kroenke et al., 1983).

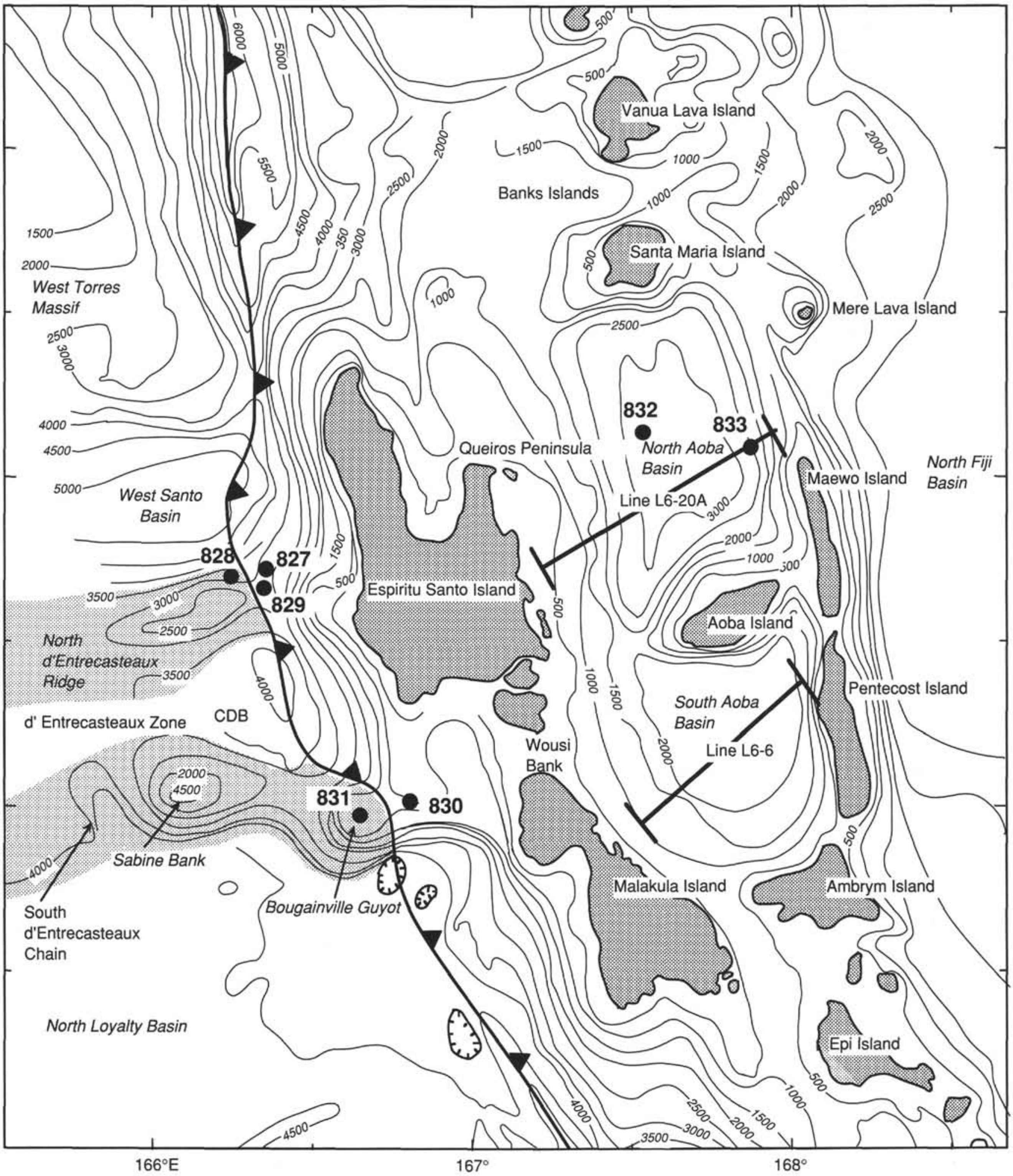

Figure 2. Detailed bathymetry (in meters) of the central New Hebrides Island Arc, showing location of Sites 827–833. Lines L6-6 and L6-20A indicate the locations of seismic profiles shown in Figures 8 and 9, respectively. CDB = Central d'Entrecasteaux Basin.

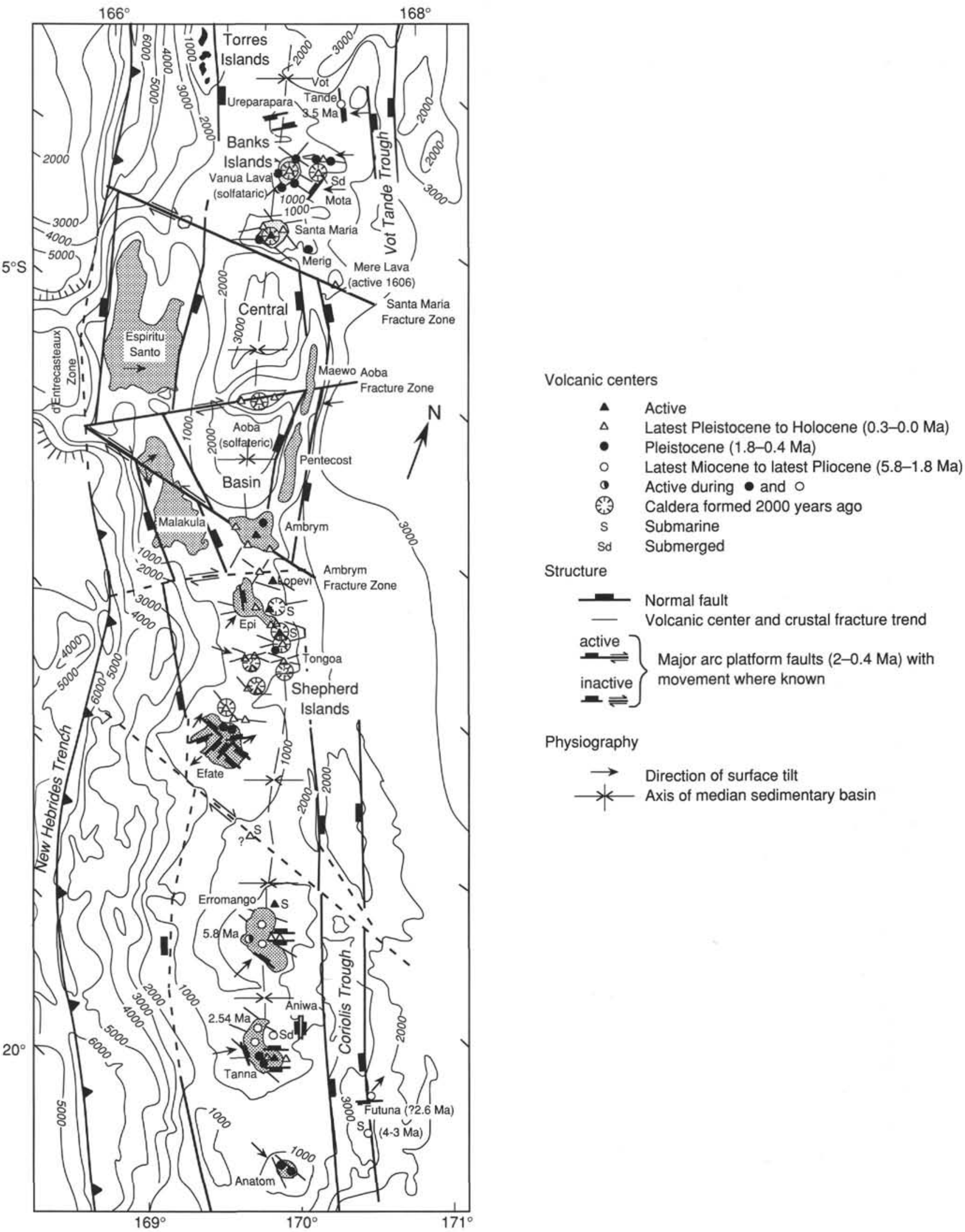

Figure 3. Tectonic map and principal volcanic centers of the New Hebrides Island Arc (after Macfarlane et al., 1988, p. 68). Bathymetry in meters.

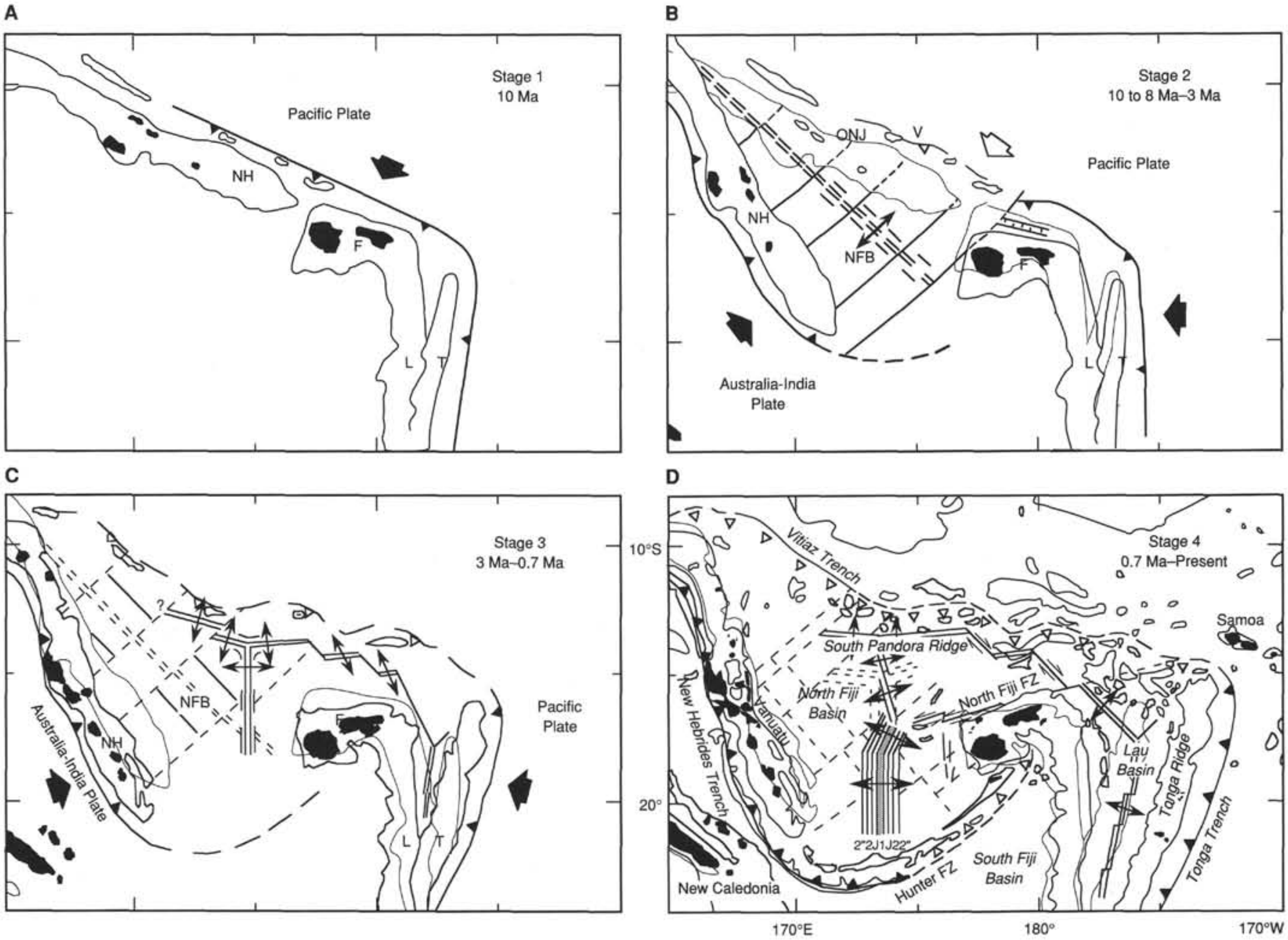

Figure 4. Four-stage evolution (A–D) of the North Fiji Basin (after Auzende et al., 1988, p. 928). Major ridges are defined as follows: New Hebrides (NH), Fiji (F), Vitiaz (V), Lau (L), and Tonga (T). Other locations defined: Ontong Java Plateau (ONJ) and North Fiji Basin (NFB). Thick black arrows indicate active subduction, white arrow (Fig. 4B, stage 2) represents fossil subduction, and thin black arrows show zones of active spreading. Thick dashed lines represent flow lines of New Hebrides rotation and thin dashed lines show magnetic lineations. The symbols "1, J, 2, and 2'" (Fig. 4D, stage 4) represent identified magnetic anomalies. FZ = fracture zone.

southwestward migration of the backarc, with the New Hebrides Island Arc moving away from Fiji, trailing a newly developing volcanic arc behind it. Magnetic anomalies and Seabeam maps of parts of the North Fiji Basin (Auzende et al., 1988) are consistent with a four-stage evolution of the North Fiji Basin since the late Miocene (Fig. 4). During stage 1, about 10 Ma, a single arc consisting of the Vitiaz zone and the New Hebrides, Fiji, Lau, and Tonga-Kermadec islands was located between the Australia-India plate and the Pacific plate (Gill and Gorton, 1973; Falvey, 1978; Malahoff et al., 1982; Fig. 4A). About 8–10 Ma (stage 2), the Ontong-Java Plateau collided with the Solomon Island Arc along the southwest-dipping Vitiaz subduction zone, the consequence of which was a possible reversal in subduction dip from southeastward along the Vitiaz arc to eastward along the New Hebrides Island Arc. This resulted in the opening of the North Fiji Basin along a spreading center, and clockwise rotation of the New Hebrides Island Arc and the counterclockwise rotation of the Fiji Islands (Fig. 4B). By about 3 Ma (stage 3), the original spreading center gave way to a north-south-trending spreading center in the central part of the North Fiji Basin, and a triple junction with a spreading axis trending east-west formed in the northern part of the basin (Fig. 4C). At 0.7 Ma (stage 4), rearrangements of the axial spreading zone and the triple junction of the North Fiji Basin occurred and the North Fiji Fracture Zone developed (Fig. 4D). These developments brought the New Hebrides Island Arc and the North Fiji Basin to their present-day configuration.

The earliest tectonic events recognized are volcanic eruptions onto the ocean floor during the late Oligocene or early Miocene, where the Western Belt islands of Espiritu Santo and Malakula are located, west of the Vitiaz frontal arc (Carney et al., 1985). Block faulting and deposition of volcanogenic sediment in adjacent flanking basins accompanied the volcanism. A compressive phase of tectonism at the beginning of the middle Miocene, possibly associated with major readjustments of the Australia-India plate, brought volcanism to a close with the emplacement of stocks into a wrench-fault system subparallel to the arc (Fig. 5A; Carney et al., 1985). A late middle Miocene tensional phase of normal faulting formed grabens, and volcanogenic sediment was deposited into the basins directly west of the main volcanic axes (Fig. 5B). At that time, the location of the Eastern Belt of islands (Maewo and Pentecost) was the site of deep-water sedimentation and deposition of terrigenous clasts derived from an uplifted and eroding Vitiaz frontal arc to the east. However, the Vitiaz frontal arc was still active during the middle Miocene (Carney et al., 1985), as recorded by the 12.7-Ma volcanism on Mitre

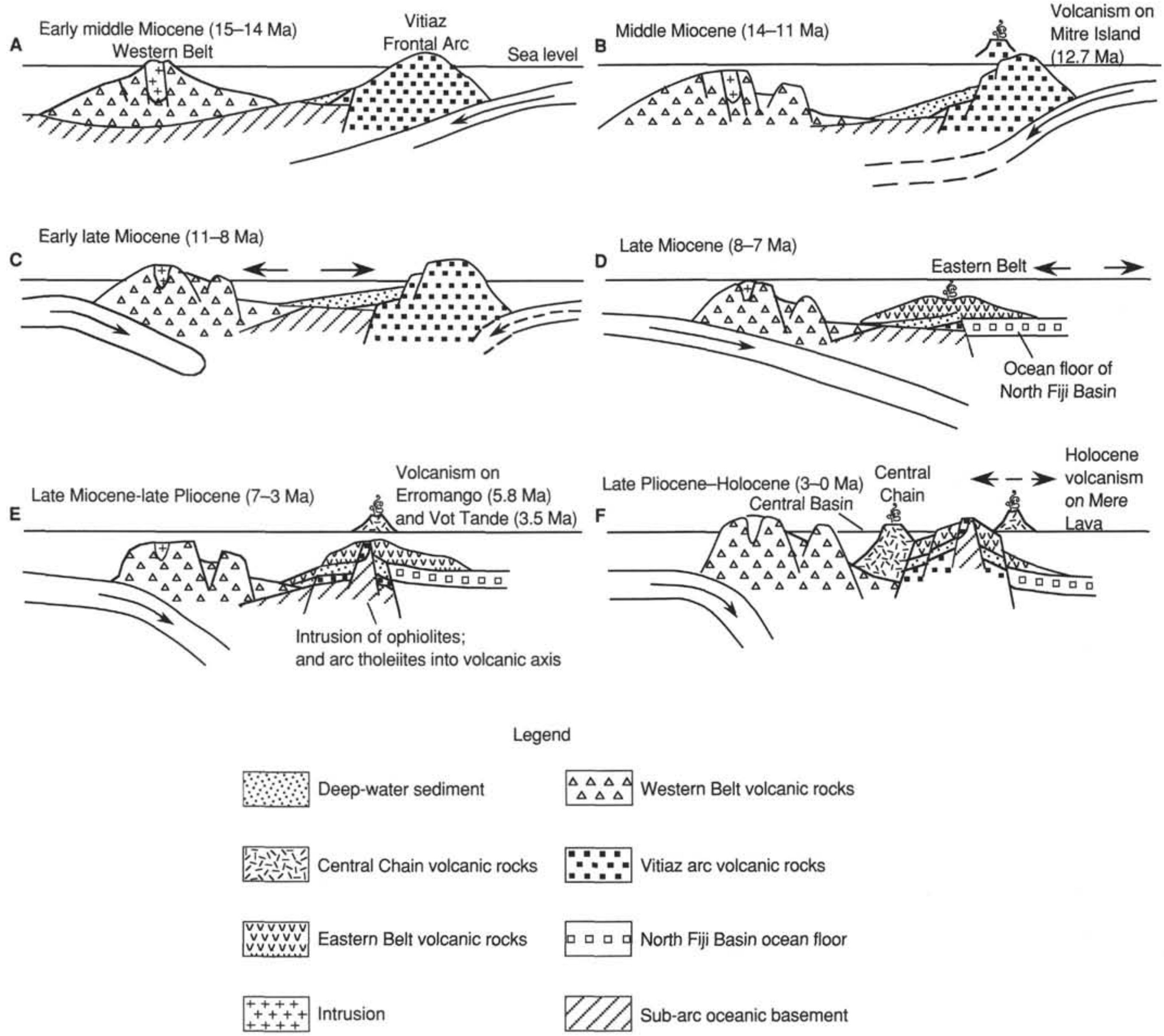

Figure 5. Model of the evolution of the New Hebrides Island Arc (A–F) (from Carney et al., 1985; Greene et al., 1988; and Macfarlane et al., 1988).

Island (Jezek et al., 1977). Initial sedimentation was, therefore, contemporaneous with tholeiitic eruptions onto the Oligocene ocean floor (Fig. 5B; Macfarlane et al., 1988).

During the early late Miocene, possible arc-polarity reversal brought about east-dipping subduction beneath the Western Belt. Strong coupling between the new downgoing plate and the new upper plate margin caused uplift of the Western Belt accompanied by folding and transverse wrench faulting; subduction at the Vitiaz Trench ceased (Fig. 5C). Extensional tectonics associated with the opening of the North Fiji Basin during the late Miocene brought about subsidence of the Western Belt and new arc volcanism at the Eastern Belt (Fig. 5D). The first collision of the proto-DEZ with the arc, which may have involved a seamount from south of Malakula, may have taken place then (Greene et al., 1988).

Partly contemporaneous with volcanism along the Eastern Belt were the earliest known eruptions along the Central Chain on Erromango during the late Miocene, and on Vot Tande and south of Futuna during the middle Pliocene (Fig. 5E). Collision of the DEZ proper with the Western Belt followed during the middle Pliocene and caused uplift of the Eastern Belt, diapiric intrusion and unroofing of the basement complex, and cessation of arc volcanism on Pentecost (Carney and Macfarlane, 1982). A further response to this localized compression was the inception of the Vot Tande and Coriolis troughs and the termination of the earliest Central Chain eruptions. Large reef-derived carbonate lenses in the pelagic sequence on Espiritu Santo Island suggest that periodic uplift interrupted the general subsidence of the Western Belt during late Miocene time (Fig. 5E; Carney and Macfarlane, 1985).

During the late Pliocene, Central Chain volcanism appears to have been distributed over a much wider belt—at least in the southern part of the arc (e.g., between Erromango, Tanna, and Futuna)—than at present; this difference argues for a more gently dipping Benioff zone at that time (Carney et al., 1985). Renewed subduction of the DEZ during the latest Pliocene caused rapid uplift of both the Western and Eastern belts during the Quaternary and formation of a deep intra-arc sedimentary basin. Volcanism at the rear of the arc-platform complex (Futuna) was terminated, the seismic zone steepened to its present inclination of 70°, and Central Chain activity, confined now to a much narrower belt than previously, spread along the length of the arc during the earliest Pleistocene to the present. However, the recently active Mere Lava volcano in the Banks Islands group (Fig. 3), which lies east of the present volcanic line, is thought to have originated from magmatism associated with incipient backarc rifting at the Vot Tande Trough (Fig. 5F; Macfarlane et al., 1988).

The northern intra-arc sedimentary basin of the New Hebrides Island Arc, although less well-developed physiographically than in the Central Basin region, can nevertheless be traced northward as a continuous feature into the eastern Solomon Islands (Katz, 1988). Its formation, therefore, cannot be linked simply to subduction of the DEZ but must relate to regional plate movements, of which the ridge-arc collision was only a part (Carney and Macfarlane, 1982; Falvey and Greene, 1988).

Proposed Origin

Many different explanations have been proposed to describe the origin of the New Hebrides Island Arc (e.g., Chase, 1971; Dickinson, 1973; Pascal et al., 1978; Falvey, 1975; Coleman and Packham, 1976; Ravenne et al., 1977; Carney and Macfarlane, 1977, 1978, 1979, 1980, 1982; Carney et al., 1985; Katz, 1988; Macfarlane et al., 1988). However, two major hypotheses are presently favored. One proposes a reversal in subduction polarity sometime between 8 and 6 Ma (Chase, 1971), the other favors a continuous eastward subduction throughout the Neogene (Luyendyk et al., 1974).

According to the polarity-reversal hypothesis, westward subduction of the Pacific plate beneath the Australia-India plate in the late Oligocene (or possibly earlier) formed the Western Belt in the rear-arc part of the older volcanic arc platform (Fig. 5). When the North Fiji Basin opened in the late Miocene to Holocene, possibly because of the collision of the Ontong-Java Plateau with the Solomon Island Arc to the north (Kroenke, 1984), the Australia-India plate started to subduct beneath the Pacific plate and then form the Eastern Belt and Central Chain islands (Chase, 1971; Carney and Macfarlane, 1978, 1980; Kroenke, 1984; Macfarlane et al., 1988).

The other hypothesis explains the formation of the three island belts by subduction of an east-dipping slab that changed inclination through time (Luyendyk et al., 1974; Carney and Macfarlane, 1977; Hanus and Vanek, 1983; Katz, 1988; Louat et al., 1988; Fig. 5). Hanus and Vanek (1983) and Louat et al. (1988) use modern seismicity data to distinguish two different east-dipping subduction slabs, one broken off from the other. Periodic cessation of subduction, with fragmentation of the downgoing lithospheric slab and shifting from a fairly steeply dipping Benioff zone to a shallower one and back again, is proposed to explain the west to east and back to west migration of the volcanic axis through time within the Central Basin region of the arc.

BACKGROUND TO SITES 827–831

The d'Entrecasteaux–New Hebrides Island Arc Collision Zone

The area of collision between the d'Entrecasteaux zone (DEZ) and the New Hebrides Island Arc results from the recent (8–2 Ma) subduction of the DEZ beneath the central part of the New Hebrides Island Arc. This collision has severely deformed the forearc and produced arc-wide fragmentation.

The DEZ separates the North Loyalty Basin, a 4.5-km-deep oceanic basin that is Eocene in age (Shipboard Scientific Party, 1975), from the West Santo Basin, a 5.5-km-deep oceanic basin of unknown age (Fig. 1). Near the central New Hebrides Trench, the DEZ consists of two east-west-trending morphologic features: the NDR and the SDC. These features are separated by the Central d'Entrecasteaux Basin (Fig. 2), which lies on the DEZ, at a depth of 3.5–4.0 km (Fig. 6; Daniel et al., 1977; Collot et al., 1985).

The NDR and the SDC differ in both morphology and lithology. The NDR is an elongated feature that rises 1.8 km above the Central d'Entrecasteaux Basin and about 3.0 km above the seafloor of the West Santo Basin (Fig. 6). The ridge, in part, consists of Paleogene mid-ocean ridge basalt (MORB) (Maillet et al., 1983) overlain by layered sedimentary rocks (Fisher et al., 1991). In contrast, the SDC comprises minor conical seamounts and two major flat-topped guyots: the Sabine Bank, which rises to within a few meters of the sea surface, and the Bougainville Guyot, whose flat summit lies below 1.0–1.5 km of water (Daniel et al., 1986) (Fig. 6). The Bougainville Guyot is an andesitic volcanic edifice topped by about 0.7 km of carbonate and lagoonal deposits, as indicated from seismic reflection, dredging, and submersible-dive data (Daniel et al., 1986). The oldest sediments recovered by dredging the guyot are mid-Eocene to mid-Oligocene in age (Collot and Fisher, 1991).

The collision and subduction of the NDR and the SDC beneath the Central New Hebrides Island Arc have created considerable deformation and tectonic erosion along the forearc. However, the NDR and the SDC have produced contrasting deformation in arc-slope rocks. From seismic reflection profiles, the NDR is being subducted with the formation of Wousi Bank, whereas subduction of the Bougainville Guyot has created only an anticline and shallow dipping faults in the forearc (Fisher et al., 1986).

The Bougainville Guyot has indented the arc slope by about 10 km, creating a 700-m-high antiform in arc-slope rocks (Fig. 6; Daniel et al., 1986; Fisher et al., 1991). Small anticlines and thrust faults that dip gently arcward also developed in the arc slope as a consequence of the collision (Fisher et al., in press). Dives conducted using the submersible *Nautile* in the DEZ-arc collision zone (Collot et al., 1989) indicate that the arc slope consists mostly of volcaniclastic rocks, along with some basaltic and andesitic lavas that were probably derived from the arc.

In contrast, the indentation caused by the collision of the NDR against the arc slope is neither as broad nor as deep as the one created by the guyot (Fig. 6). The most prominent morphologic feature in this collision zone is the Wousi Bank, a broad, shallow (5 m below sea level) protrusion that developed across the arc slope. The Wousi Bank is aligned with the NDR, suggesting that the bank was uplifted as a result of collision between the NDR and the arc. Drilling both the NDR and the Bougainville Guyot determined the relationship between the style of forearc deformation and the nature and physical properties of the impinging rocks.

BACKGROUND TO SITES 832–833

The Intra-Arc Basins

Geology of the Central Basin Region

The central New Hebrides Island Arc is a product of multiple phases of arc evolution (Mallick, 1975; Carney and Macfarlane, 1977; Carney et al., 1985). Stratigraphic se-

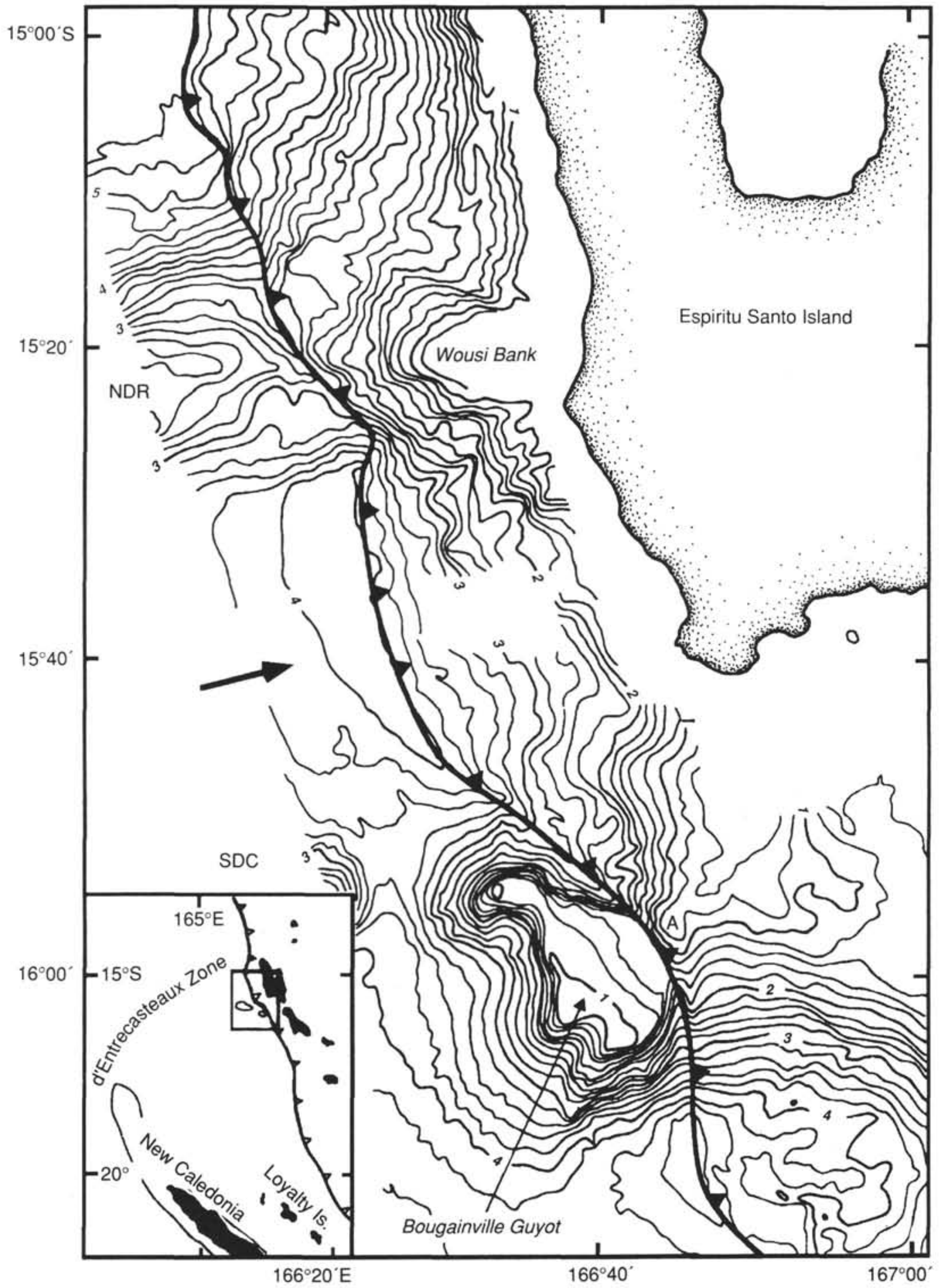

Figure 6. Seabeam bathymetry (in kilometers) of the collision zone between the d'Entrecasteaux Zone and the New Hebrides Island Arc (after Daniel et al., 1986). The symbol "A" opposite the Bougainville Guyot denotes the antiform in arc-slope rocks caused by the collision. SDC = South d'Entrecasteaux Chain; NDR = North d'Entrecasteaux Ridge.

quences of the intra-arc region confirm a complex geologic and sedimentologic history. Vertical tectonism elevated and depressed local horsts and grabens, as demonstrated by the presence of erosional and buttress unconformities. The main graben under the Central Basin is divided by transverse faulting and extrusion of volcanic rocks at Aoba and Ambrym islands. This main graben appears to have remained below sea level through most of the Neogene and Quaternary. However, regional unconformities identified in seismic reflection profiles record at least three major uplifts and erosional cycles along its eastern and western margins since the middle Miocene. At the end of uplift during the middle Miocene, folding and erosion occurred on the major horst blocks of Espiritu Santo and Malakula islands, which were succeeded by a north-to-south transgression. A younger unconformity represents a Pleistocene uplift that is also recorded on the islands by elevated marine terraces and reefs. In the Eastern Belt, deep-water sedimentation and arc volcanism prevailed between the middle Miocene and Pliocene.

In addition to the major horsts (Espiritu Santo, Malakula, Maewo, and Pentecost islands) and the major graben (North and South Aoba basins), smaller subsidiary fault blocks rose and subsided, alternately allowing erosion and deposition (Fig. 5). Big Bay and Cumberland basins of Espiritu Santo Island (Fig. 7) are both grabens, but erosion is taking place in the Big Bay Basin and deposition in the Cumberland Basin. The Queiros Peninsula is a horst overlain by deep-water sediment. Along the eastern margins of Malakula and Espiritu Santo, fault blocks define Malakula and East Santo basins (Fig. 7). These blocks are down-dropped to the east and are separated from North and South Aoba basins by a fault-controlled structural ridge; on the East Santo shelf, the ridge is capped by reefs. These blocks are inferred to be grabens or half grabens that accumulated a thick sedimentary sequence (~2 km). The Malakula Basin appears to be a half graben because it is bounded by a fault on its eastern side, and the basin fill is progressively more inclined toward the west with depth (Fig. 8). Even the modern depositional surface is inclined at 45 m/km (Johnson and Greene, 1988). The East Santo Basin is more likely to be a subsiding step block; the entire basin fill is almost horizontally layered and the present seabed has a lower gradient (25 m/km).

These sedimentary basins are structurally complex and result from island-arc tectonics that appear unique to the central Vanuatu region (Fig. 3). The formation of these basins in the Central Basin region is influenced by the collision of the DEZ with the central New Hebrides Island Arc opposite Espiritu Santo and Malakula islands (Karig and Mammerickx, 1972; Daniel and Katz, 1981; Burne et al., 1988; Greene et al., 1988; Collot et al., 1985). Since the Pliocene, this collision has been characterized by wrench-fault tectonics, that is, crustal extension and transtension creating horst-and-graben blocks, and crustal compression and transpression uplifting and shifting those same structures (Greene et al., 1988).

Distinct regions of compression and extension have been identified in the Central Basin (Fig. 3; Greene et al., 1988). For example, the Aoba Fracture Zone separates two quadrants or structural wedges. Greene et al. (1988) have speculated that before the development of the modern tectonic regime and the impingement of the DEZ upon Espiritu Santo Island, the Malakula and East Santo basins were continuous and the North and South Aoba basins were one basin. Collision, possibly during the Pliocene, initiated both right-lateral motion and magmatic extrusion along the Aoba Fracture Zone. This motion has offset the basins by as much as 25 km (Greene et al., 1988). Older sets of faults are oriented parallel to the arc (north-northwest by south-

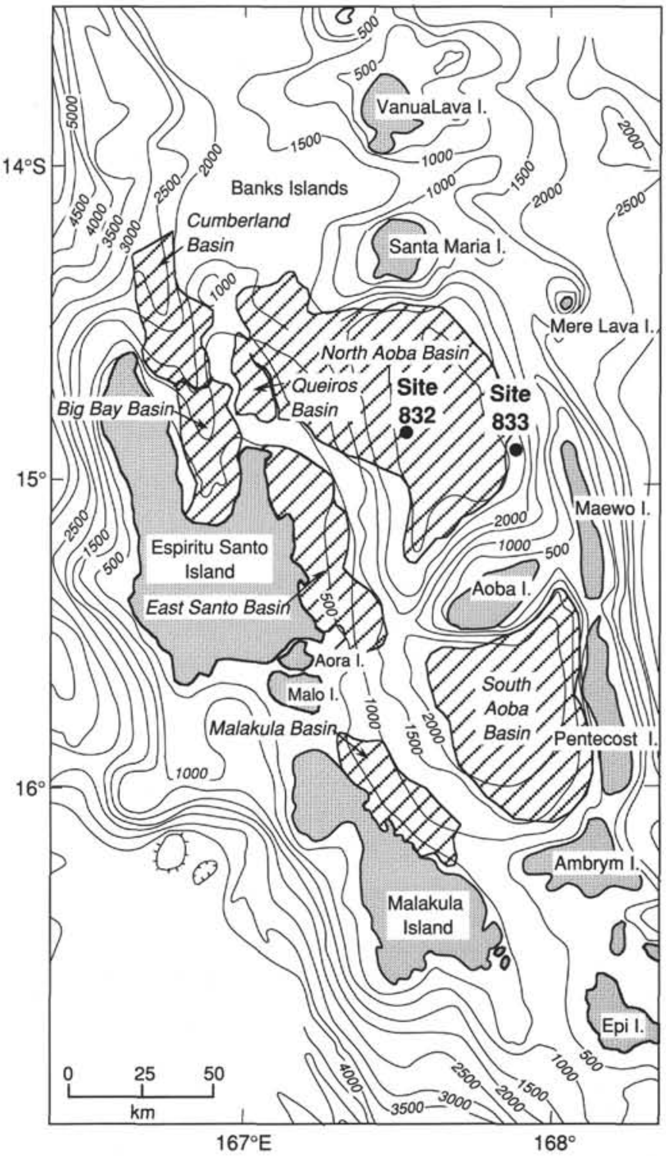

Figure 7. Sedimentary (structural) basins of the central New Hebrides Island Arc (Vanuatu) summit platform area. Modified after Greene and Johnson (1988). Bathymetry in meters.

southeast) and are composed of the steep normal faults that bound the east and west margins of the Central Basin and front the arc; these show displacement primarily downward on the basin and trench side. These faults probably formed by thermal upbowing. The faults oblique to the arc (oriented northwest-southeast) are normal faults that show recent movement, some with strike-slip sense (Greene et al., 1988). The primary faults or rifts transverse to the arc display a more random orientation (with a general east-west trend) and are transcurrent faults that result from the collision of the DEZ. Magma leaking along the northernmost rift (the Santa Maria Fracture Zone; Fig. 3) has produced the small volcanic island of Mere Lava (last recorded active in 1606), along the central rift (Aoba Fracture Zone) the large shield volcano of Aoba (active in the 1800s), and along the southern rift (Ambrym Fracture Zone) the large active shield volcano of Ambrym (Greene et al., 1988; Wong and Greene, 1988).

North Aoba Basin is a physiographic and sedimentary basin that lies between Espiritu Santo, Aoba, Maewo, and

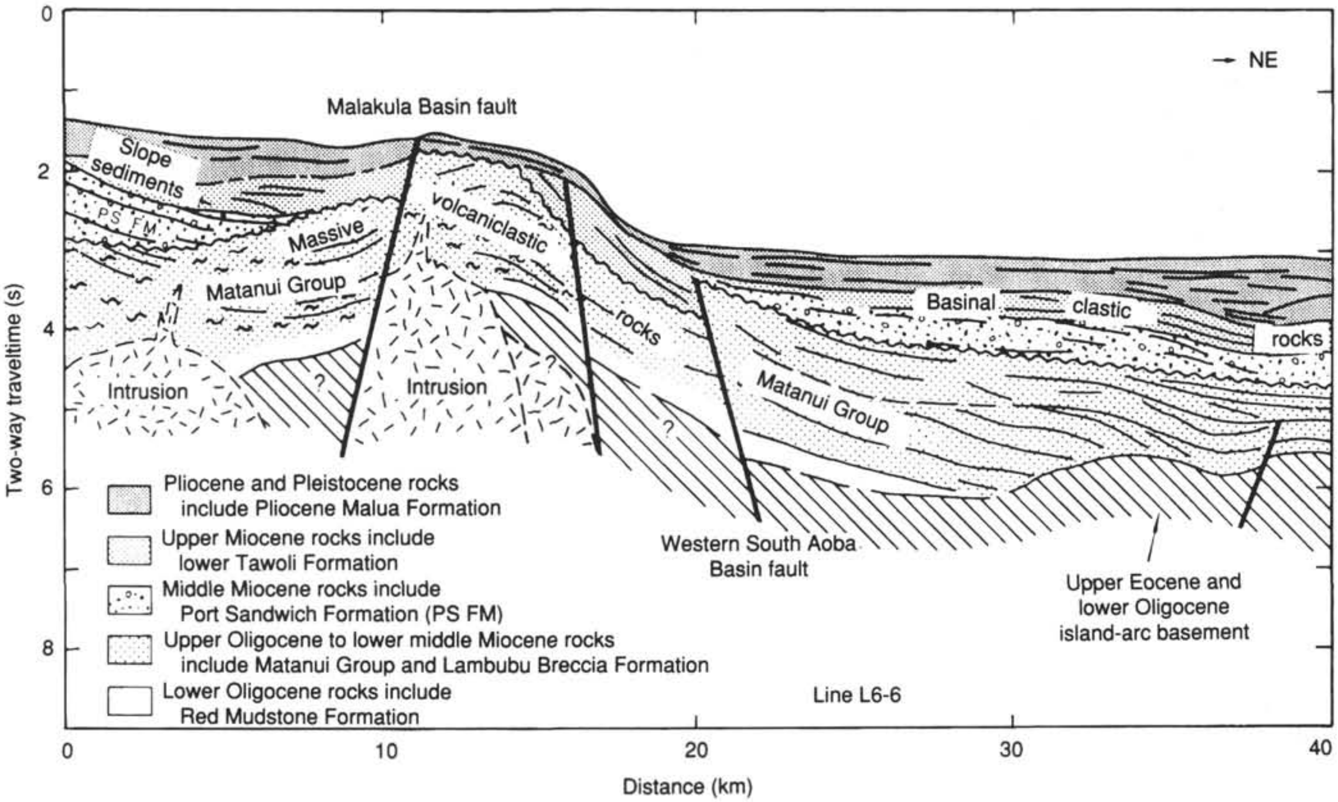

Figure 8. Interpreted multichannel seismic-reflection profile, Malakula Basin (after Falvey and Greene, 1988, p. 427). See Figure 2 for location.

Santa Maria islands (Figs. 2 and 9). The basin initially developed as a half graben tilted to the east (Katz, 1988; Fisher et al., 1988) and seismic refraction data indicate that it is filled with 5 km of sediment (Holmes, 1988). The oldest rocks identified within the basin from seismic reflection data were inferred to be lower to middle Miocene fine-grained volcaniclastic graywacke and limestone of indeterminate thickness (Greene and Johnson, 1988; Fisher et al., 1988). This sediment underlies the east slope of the basin. Unconformably overlying these rocks may be more than 1.8 km of probable upper Miocene calcarenites (Greene and Johnson, 1988; Fisher et al., 1988) that are gently faulted and lap up onto the east and west flanks of the basin. Another 0.8 km or more of uppermost Miocene to lowermost Pleistocene calcarenite and calcilutite appear from seismic reflection data to overlie the middle Miocene sedimentary rocks. These rocks, in turn, appear to have been overlain by 0.8 km of Holocene ash and pelagic sediments that are found in the flat floor of the basin.

East Santo Basin is a small, linear island-shelf basin elevated above the west flank of the North Aoba Basin. East Santo Basin trends north-south parallel to the east coast of Espiritu Santo and, with the exception of its eastern structural boundary, has no physiographic expression (Fig. 7). The western boundary is not well defined, but the eastern boundary of Malakula Basin is a structural and bathymetric ridge that appears to have been folded into an anticline and possibly faulted by a high-angle fault that is capped with carbonate rocks. This structure is sinuous and extends for over 70 km in a general north-south direction. The basin is truncated in the south by faults or overlain with lava flows from Aoba. Previous work (Katz, 1988) indicates that the basin has resulted from the subsidence of a block bounded by a reef-capped faulted and folded ridge (Greene et al., 1988; Greene and Johnson, 1988). The basin appears to have formed by the infilling of this slowly down-dropping reef-bounded block.

Seismic reflection data indicate that East Santo Basin is filled with nearly 2 km of sediment, the lower part consisting of upper Miocene to Pliocene biocalcarenite (correlated with onshore deposits) more than 1 km thick, and the upper part consisting of about 1 km of Miocene to lower Pliocene interbedded calcarenite, calcilutite, and foraminiferal mudstone (Greene and Johnson, 1988; Fisher et al., 1988). Falvey and Greene (1988) suggest that these deposits were ponded behind (west of) the structural and bathymetric ridge that marks the basin's eastern boundary because the strata dip eastward, indicating that they were derived from the island of Espiritu Santo to the west. The deepest part of the basin is adjacent to the structural high where the basin axis extends in a general north-south direction (Fig. 3).

Regional Evolution of the Intra-Arc Basin

Four major phases (I–IV) of sedimentation and arc volcanism, separated by regional unconformities, have contributed to the evolution of the sedimentary sequences in the present intra-arc basin. The three major unconformities are recognized on the surrounding islands, and seismic reflection profiles (Greene and Johnson, 1988; Fisher et al., 1988) show these unconformities to extend into and across the intra-arc basin, grading into disconformable or even conformable contacts in the central part of the basin. The lowest unconformity can be clearly identified only on multichannel profiles and is interpreted to be early middle Miocene in age (Fisher et al., 1988). The middle and upper unconformities can be identified on both single-channel and multichannel profiles (Fisher et al., 1988); one was interpreted to be of middle Miocene age, and

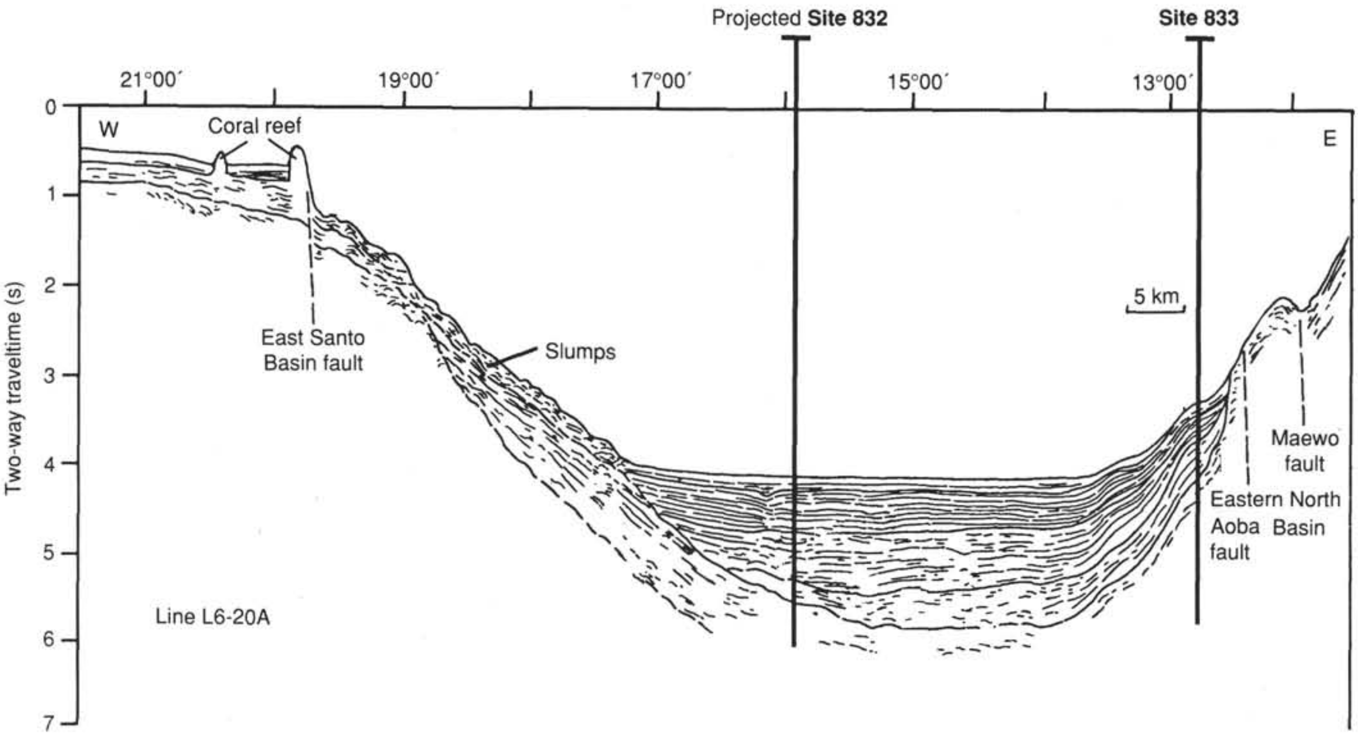

Figure 9. Interpreted seismic-reflection profile across the East Santo and North Aoba basins (after Greene and Johnson, 1988, p. 187). See Figure 2 for location.

the others were interpreted as late Pliocene or early Quaternary.

Phase I: Oligocene to Early Middle Miocene

Sequences interpreted as Oligocene to lower middle Miocene are the lowest units observable on seismic reflection profiles (Greene and Johnson, 1988), and are not traceable under the whole arc platform. Onshore evidence of the geological setting at this time is also fragmentary. It is thought that arc volcanism, associated with a southwest-dipping subduction zone near the present Vitiaz Trench, was established by late Oligocene time and perhaps as early as late Eocene (Mallick, 1975; Kroenke, 1984). Mainly massive submarine arc volcanic rocks (lava, breccia, and sedimentary sequences) were deposited along the Western Belt (e.g., Mitchell, 1970). Interbedded limestone indicates that shallow-water conditions generally prevailed. The sequences were intruded by lower to middle Miocene andesitic dikes and sills. Along the Eastern Belt, minor amounts of deep-water sediments were deposited over oceanic crust (Mallick and Neef, 1974; Macfarlane et al., 1988).

Phase I volcanism was concentrated in the Western Belt and appears to have lasted intermittently for at least 11 m.y. (Kroenke, 1984; Greene et al., 1988; Macfarlane et al., 1988). It was followed by block faulting, uplift, and erosion.

Phase II: Middle Miocene

Above the lower middle Miocene unconformity is a seismic unit composed of irregular and discontinuous reflectors that can be traced under much of the intra-arc basin. In the Western Belt, basinal graywacke-filled, block-faulted grabens at Espiritu Santo, and shallow-water andesitic graywacke, tuff, and limestone accumulated at Malakula. In the Eastern Belt, the sedimentary sequence passes upward from pelagic mudstone to sandstone to limestone and conglomerate, indicating shallowing and a nearby source terrain. Detailed work in the lower part of the Sighotara Group on Maewo (Neef et al., 1985) indicates deposition of a deep-sea fan below the carbonate compensation depth in an intra-arc basin. Although both the lower middle Miocene and middle Miocene unconformities are widely recognized within the Western Belt on land, their distribution offshore is less clear. Both are recognized in the Malakula Basin, but neither is apparent in the East Santo Basin. However, these two unconformities converge from the east (Central Basin region) over the structural high that marks the eastern side of both basins. In the North and South Aoba basins, both unconformities are recognized, although the middle Miocene unconformity is less pronounced than the earlier one. However, in the Eastern Belt the middle Miocene unconformity appears to be absent.

Seismic reflection profiles reveal a much thicker sequence between the two unconformities under the central and eastern parts of the North and South Aoba basins than is found onshore (Greene and Johnson, 1988). Two-way traveltime in the submarine sequence is at least one second, equivalent to perhaps 1.3 km, assuming a velocity of 2.5 km/s. The section exposed onshore, however, is less than 1 km thick.

Phase III: Late Miocene to Pliocene

Two discontinuous seismic reflection sequences, separated by an unconformity, lie above the middle Miocene unconformity and can be traced across the entire intra-arc basin (Greene and Johnson, 1988). These sequences and the unconformity between them indicate that initial patterns of sedimentation varied across individual basins. In the Malakula and East Santo basins the seismic reflection sequences are nearly flat-lying and appear to be ponded behind a structural ridge. These structural trends, generally north-south to north-northwest-south-southeast, must have begun to develop as early as the late Miocene (Greene et al., 1988). In the central parts of the North and South Aoba basins, the seismic reflection sequences exhibit their thickest development and show prominent bedding with extensive lateral continuity, indicating continuous deposition on a broad, slowly subsiding platform. To the west, strata progressively onlap the surface of the middle Miocene seismic reflection sequence (Fig. 8), demonstrating that the Western Belt was then an emerging topographic high. To the east, upper Miocene to Pliocene beds rise sharply and are draped along the flanks of the Eastern Belt islands. Thus, in contrast to the Western Belt, the main uplift of the Eastern Belt must postdate the late Miocene to Pliocene.

Onshore geologic relations confirm the initial emergence of the Western Belt islands. On Malakula Island, deposition was delayed until the basal Pliocene transgression deposited an upward-deepening sedimentary sequence (Malua and Wintua formations). On Espiritu Santo Island, however, deposition of calcarenite took place from the early late Miocene until it was replaced by deeper-water mudstone in the Pliocene.

In summary, the upper Miocene to Pliocene sequences record overall subsidence of the central and eastern region of the arc, with the western part also subsiding in the Pliocene. Arc volcanism was limited to the southern half of Maewo Island and throughout Pentecost Island in the Eastern Belt and lasted perhaps 4 m.y. (from 7 to 3 Ma).

Phase IV: Quaternary

The margins of the arc were uplifted in the early Quaternary. This vertical tectonism was considerable; upper Pliocene pelagic mudstone, which must have been deposited in at least several hundred meters of water, was raised hundreds of meters above sea level. On the seismic reflection profiles, Holocene strata appear to be as much as 0.5 s (perhaps 375 m) thick in the central North Aoba Basin, but thin laterally and are draped over the basin's margins. Downslope mass movements are particularly evident along the western margin. These offshore features confirm the major differential subsidence of the intra-arc basin compared to the rising Eastern and Western belts of islands. Pelagic volcanic ash and skeletal carbonate accumulated in the basin (Johnson and Greene, 1988). Reefal limestone, which formed around the islands and unconformably overlies earlier units, has been progressively uplifted. Uplift rates have been estimated at 1.7–3.1 m/10^3 yr on Pentecost (Mallick and Neef, 1974), as much as 3.5 m/10^3 yr on Malakula and 5 m/10^3 yr on Espiritu Santo (Jouannic et al., 1980; Taylor et al., 1985). Volcanic eruptions along the Central Chain accompanied this Quaternary uplift. This volcanism, directly following the earlier Miocene to Pliocene activity in the Eastern Belt, indicates continuing subduction. However, a major change, steepening of the subduction zone, is indicated by (1) the westward shift in the volcanic arc and (2) the dramatic expansion of volcanism from a limited area in the Eastern Belt to an extensive Central Chain (Macfarlane et al., 1988).

The South Aoba Basin is surrounded by the islands of Aoba, Pentecost, Ambrym, and Malakula (Fig. 2). This physiographic and sedimentary basin is a graben with well-developed faults on its east and west margins. The basin is filled with more than 3 km of poorly layered sedimentary rocks (Fisher et al., 1988). The island-shelf Malakula Basin has no physiographic expression and lies beneath the relatively broad eastern shelf of Malakula Island (Fig. 8). The basin's eastern margin is a structural bedrock ridge over 50 km long that has ponded sediments behind it to the west. This somewhat sinuous structural ridge is a broad anticline faulted into a horst that is buried beneath a thin cover of Quaternary marine sediments (Greene and Johnson, 1988). The northern and southern limits of this ridge are truncated by the Aoba and Ambrym fracture zones or are covered by submarine lava flows from these volcanoes.

Sedimentation in the Deep Intra-Arc Basins

The two deep basins (North and South Aoba) are enclosed depressions with relatively flat floors that comprise most of the area of the Central Basin region. They are separated by an east-west-trending volcanic ridge surmounted by Aoba Island. Before the early Quaternary, a single elongate northward-deepening basin (the Central Basin of Ravenne et al., 1977) extended over 150 km north-south, which was later subdivided by an emerging volcano (Greene et al., 1988). The northern deep basin is approximately rectangular—35 km wide (east-west) by 55 km long (north-south) and over 3 km deep. The southern deep basin is squarer in outline, approximately 35 km across, and has a floor that slopes gently from 2.0 km in the south to 2.4 km in the north. The basins are surrounded by sills that appear to be unbreached below 1.8 km of water depth. The physiographic basins are largely Quaternary features that overlie older basin fill (Carney and Macfarlane, 1980; Fisher et al., 1988; Greene and Johnson, 1988). Seismic reflection profiles indicate that the deep basins contain flat, well-layered sediments in which individual reflectors can be traced basin-wide (Fisher et al., 1988; Greene and Johnson, 1988).

Mudstones dredged off Malakula Island and Malo Island (just south of Espiritu Santo Island) range in age from early middle Pliocene to late Pleistocene or Holocene (Johnson and Greene, 1988). Calcarenite, recrystallized limestones, and coral samples collected offshore are correlated with Quaternary reefal limestones, which form terraces onshore (Johnson and Greene, 1988). Samples of limestone taken from resistant substrate, and the few reefal limestone samples recovered from relatively deep water (~500 m), indicate considerable Quaternary subsidence along the western side of the Central Basin. Uplift of the islands is well documented from onshore studies of raised reef terraces (Jouannic et al., 1980; Taylor et al., 1980).

Suites of volcanic sandstone and mudstone, generally composed of vitric and crystal ash with scattered foraminifers and other skeletal debris, were dredged off the eastern margin of North and South Aoba basins. Foraminifers indicate ages from middle late Miocene to Pliocene, with some ranging into the Pleistocene. These lithologies are correlated with the upper Miocene to Pliocene formations on Maewo Island and equivalent units on Pentecost Island (Johnson et al., 1988). However, two samples were of solely Pleistocene age, indicating that similar sedimentation continued offshore well after the upper Tertiary units had been uplifted. The Pleistocene samples may equate with the sandstone and mudstone of the late Pliocene and early Pleistocene in northern Maewo Island. They are also very similar to the lithologies recovered from the western margin of the North and South Aoba basins, which have been correlated with units of similar age onshore.

One dredge haul between Pentecost and Maewo islands recovered vesicular basaltic andesite compositionally similar to upper Miocene to Pliocene volcanic rocks on Maewo and Pentecost (Johnson et al., 1988). These onshore volcanic rocks contain basalt, basaltic andesite, and andesite with about 0.8%–0.9% TiO_2, and K/Rb ratios for the basaltic andesite ranging from 500 to 700, with an average of 600. The K/Rb ratio for the dredge sample is 575. Comparisons of rare-earth element data support this correlation and the sample shows a slight europium anomaly reflecting plagioclase accumulation.

DRILLING OBJECTIVES

Arc-Ridge Collision Sites

Sites within the collision zone were designed to determine the influence that ridge composition and structure exert on the style of accretion and type of arc structures produced during collision. Sites 827 and 829 are located where the northern ridge (NDR) of the DEZ and the arc collide (Fig. 2). At Site 828 we sought to obtain a critical reference section of NDR rocks to enable recognition of these rocks in other drill holes (Fig. 2). At Sites 827 and 829, we sought to penetrate the lowermost accretionary wedge, the interplate thrust fault (décollement), and the NDR itself. We hoped that this site would show whether rocks of the northern ridge have been accreted onto the arc, as well as reveal the age and mechanical properties of rocks where, despite the great relief of the subducted ridge, the collision has apparently caused little indentation and shortening in the forearc.

Sites 830 and 831 are located where the Bougainville Guyot has collided with the arc, causing considerable indentation and shortening in the forearc (Figs. 2 and 6). At Site 830 we sought to penetrate imbricated arc rocks to test whether these rocks are part of an uplifted old accretionary wedge, recently accreted guyot rocks, or island-arc basement. At Site 831 our objective was to determine the lithology, age, paleobathymetry, and mechanical properties of the guyot. Results obtained from drilling near the guyot were contrasted with those obtained near the north ridge to determine why arc structures induced by the collision are so different. The rate of uplift of the accretionary wedge will be determined and compared to the rate at which onshore areas emerged; this emergence occurred synchronously with collision, and onshore areas rose at Holocene rates exceeding 5 mm/yr.

Intra-Arc Basins

Sites 832 and 833 in the North Aoba Basin (Fig. 2) were drilled to investigate how arc-ridge collision affected the development of the intra-arc basins and the evolution of the magmatic arc. In addition, we sought to determine if volcanic ashes within basin rocks contain a record of the hypothesized reversal in arc polarity.

Site 832 is located within the center of the North Aoba Basin (Fig. 2), which contains significantly deeper water than does any other basin near the summit of this arc. Our objective was to obtain the age of a major unconformity that likely correlates with the onset of arc-ridge collision and should provide one of the better estimates of when this onset occurred. Chemistry of Quaternary volcanic ashes will be used to show if the magmatic arc has been affected by subduction of the DEZ.

Site 833 is located along the eastern flank of the North Aoba Basin (Fig. 2), where basin rocks include two unconformities. The shallower one provides temporal constraints on backarc deformation, possibly as a direct result of the collision. The deeper unconformity lies along the top of the oldest basin rocks and drilling will help to elucidate the late Cenozoic evolution of the magmatic arc. The chemistry of volcanic ash should show whether the magmatic arc was affected by the arc polarity change.

REFERENCES

Auzende, J. M., Lafoy, Y., and Marsset, B., 1988. Recent geodynamic evolution of the north Fiji basin (Southwest Pacific). *Geology,* 16:925–929.

Berggren, W. A., Kent, D. V., Flynn, J. J., and Van Couvering, J. A., 1985. Cenozoic geochronology. *Geol. Soc. Am. Bull.,* 96:1407–1418.

Burne, R. V., Collot, J.-Y., and Daniel, J., 1988. Superficial structures and stress regimes of the downgoing plate associated with subduction-collision in the Central New Hebrides Arc (Vanuatu). *In* Greene, H. G., and Wong, F. L. (Eds.), *Geology and Offshore Resources of Pacific Island Arcs—Vanuatu Region.* Circum-Pac. Council Energy Miner. Resour., Earth Sci. Ser., 8:357–376.

Carney, J. N., and Macfarlane, A., 1977. Volcano-tectonic events and pre-Pliocene crustal extension in the New Hebrides. *Int. Symp. on Geodyn. in South-west Pac., Noumea, New Caledonia, 1976.* Paris (Editions Technip), 91–104.

_______, 1978. Lower to middle Miocene sediments on Maewo, New Hebrides, and their relevance to the development of the Outer Melanesian Arc system. *Australas. Soc. Explor. Geophys. Bull.* 9:123–130.

_______, 1979. Geology of Tanna, Aneityum, Futana and Aniwa. *Regional Rept.—New Hebrides Geol. Surv.*

_______, 1980. A sedimentary basin in the central New Hebrides Arc. *Tech. Bull.—U. N. Econ. Soc. Comm. Asia Pac., Comm. Co-ord Jt. Prospect Miner. Resour. South Pac. Offshore Areas,* 3:109–120.

_______, 1982. Geological evidence bearing on the Miocene to Recent structural evolution of the New Hebrides Arc. *Tectonophysics,* 87:147–175.

_______, 1985. Geology and mineralisation of the Cumberland Peninsula, north Espiritu Santo. *Vanuatu Dept. of Geol., Mines and Rural Water Suppl. Rep.*

Carney, J. N., Macfarlane, A., and Mallick, D.I.J., 1985. The Vanuatu island arc—an outline of the stratigraphy, structure, and petrology. *In* Nairn, A.E.M., Stehli, F. G., and Uyeda, S. (Eds.), *The Ocean Basins and Margins* (Vol. 7): New York (Plenum), 685–718.

Chase, C. G., 1971. Tectonic history of the Fiji Plateau. *Geol. Soc. Am. Bull.,* 82:3087–3110.

Coleman, P. J., and Packham, G. H., 1976. The Melanesian borderlands and India-Pacific plate boundaries. *Earth-Sci. Rev.,* 12:197–233.

Collot, J.-Y., and Fisher, M. A., 1991. The collision zone between the North d'Entrecasteaux Ridge and the New Hebrides Island Arc. Part 1: Seabeam morphology and shallow structure. *J. Geophys. Res.,* 96:4457–4478.

Collot, J.-Y., Daniel, J., and Burne, R. V., 1985. Recent tectonics associated with the subduction/collision of the D'Entrecasteaux zone in the central New Hebrides. *Tectonophysics,* 112:325–356.

Collot, J.-Y., Pelletier, B., Boulin, J., Daniel, J., Eissen, J.-P., Fisher, M. A., Greene, H. G., Lallemand, S., and Monzier, M., 1989. Premiers résultats des plongées de la campagne SUBPSO1 dans la zone de collision des rides d'Entrecasteaux et de l'arc des Nouvelles-Hébrides. *C. R. Acad. Sci. Ser. 2,* 309:1947–1954.

Daniel, J., 1978. Morphology and structure of the southern part of the New Hebrides island arc system. *J. Phys. Earth,* 26:S181–S190.

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., Collot, J. Y., Monzier, M., Pelletier, B., Butscher, J., Deplus, C., Dubois, J., Gérard, M., Maillet, P., Monjaret, M. C., Récy, J., Renard, V., Rigolot, P., and Temakon, S. J., 1986. Subduction et collision le long de l'arc des Nouvelles-Hébrides (Vanuatu): résultats préliminaires de la campagne SEAPSO (leg 1). *C. R. Acad. Sci. Ser. 2,* 303:805–810.

Daniel, J., Jouannic, C., Larue, B., and Récy, J., 1977. Interpretation of d'Entrecasteaux zone (north of New Caledonia). *Int. Symp. on Geodyn. in South-west Pac., Noumea, New Caledonia, 1976.* Paris (Editions Technip), 117–124.

Dickinson, W .R., 1973. Widths of modern arc-trench gaps proportional to past duration of igneous activity in associated magmatic arcs. *J. Geophys. Res.,* 78:3376–3389.

Falvey, D. A., 1975. Arc reversals and a tectonic model for the North Fiji Basin. *Austral. Soc. Explor. Geophys. Bull.,* 6:47–49.

_______, 1978. Analysis of paleomagnetic data from the New Hebrides. *Austral. Soc. Explor. Geophys. Bull.,* 9:117–123.

Falvey, D. A., and Greene, H. G., 1988. Origin and evolution of the sedimentary basins of the New Hebrides Arc. *In* Greene, H. G., and Wong, F. L. (Eds.), *Geology and Offshore Resources of Pacific Island Arcs—Vanuatu Region.* Circum-Pac. Counc. Energy and Miner. Resour., Earth Sci. Ser., 8:413–442.

Fisher, M. A., 1986. Tectonic processes at the collision of the D'Entrecasteaux zone and the New Hebrides island arc. *J. Geophys. Res.*, 91:10470–10486.

Fisher, M. A., Collot, J.-Y., and Geist, E. L., 1991. The collision zone between the North d'Entrecasteaux Ridge and the New Hebrides Island Arc. Part 2: structure from multichannel seismic data. *J. Geophys. Res.*, 96:4479–4495.

———, in press. Structure of the collision zone between Bougainville Guyot and the accretionary wedge of the New Hebrides Island Arc, Southwest Pacific. *Tectonics*.

Fisher, M. A., Collot, J.-Y., and Smith, G. L., 1986. Possible causes for structural variation where the New Hebrides island arc and the D'Entrecasteaux zone collide. *Geology*, 14:51–954.

Fisher, M. A., Falvey, D. A., and Smith, G. L., 1988. Seismic stratigraphy of the summit basins of the New Hebrides Island Arc. *In* Greene, H. G., and Wong, F. L. (Eds.), *Geology and Offshore Resources of Pacific Island Arcs—Vanuatu Region*. Circum-Pac. Counc. Energy and Miner. Resour., Earth Sci. Ser., 8:201–224.

Gill, J. B., and Gorton, M. P., 1973. A proposed geological and geochemical history of eastern Melanesia. *In* Coleman, P. J. (Ed.), *The Western Pacific: Island Arcs, Marginal Seas and Geochemistry:* Perth (Univ. of Western Australia Press), 543–566.

Greene, H. G., and Johnson, D. P., 1988. Geology of the Central Basin region of the New Hebrides Arc inferred from single-channel seismic-reflection data. *In* Greene, H. G., and Wong, F. L. (Eds.), *Geology and Offshore Resources of Pacific Island Arcs—Vanuatu Region*. Circum-Pac. Counc. Energy and Miner. Resour., Earth Sci. Ser., 8:177–200.

Greene, H. G., and Wong, F. L. (Eds.), 1988. *Geology and Offshore Resources of Pacific Island Arcs—Vanuatu Region*. Circum-Pac. Counc. Energy and Miner. Resour., Earth Sci. Ser., No. 8.

Greene, H. G., Macfarlane, A., Johnson, D. A., and Crawford, A. J., 1988. Structure and tectonics of the central New Hebrides Arc. *In* Greene, H. G., and Wong, F. L. (Eds.), *Geology and Offshore Resources of Pacific Island Arcs—Vanuatu Region*. Circum-Pac. Counc. Energy and Miner. Resour., Earth Sci. Ser., 8:377–412.

Hanus, V., and Vanek, J., 1983. Deep structure of the Vanuatu (New Hebrides) islands arc: intermediate depth collision of subducted lithospheric plates. *N. Z. J. Geol. Geophys.*, 26:133–154.

Holmes, M. L., 1988. Seismic refraction measurements in the summit basins of the New Hebrides Arc. *In* Greene, H. G., and Wong, F. L. (Eds.), *Geology and Offshore Resources of Pacific Island Arcs—Vanuatu Region*. Circum-Pac. Counc. Energy and Miner. Resour., Earth Sci. Ser., 8:163–176.

Isacks, B. L., Cardwell, R. K., Chatelain, J. L., Barazangi, M., Marthelot, J.-M., Chinn, D., and Louat, R., 1981. Seismicity and tectonics of the central New Hebrides island arc. *In* Simpson, D. W., and Richards, P. G. (Eds.), *Earthquake Prediction: An International Review*. Am. Geophys. Union, Maurice Ewing Ser., 4:93–116.

Jezek, P. A., Bryan, W. B., Haggerty, S. E., and Johnson, H. P., 1977. Petrography, petrology, and tectonic implications of Mitre Island, northern Fiji Plateau. *Mar. Geol.*, 24:123–148.

Johnson, D. P., and Greene, H. G., 1988. Modern depositional regimes, offshore Vanuatu. *In* Greene, H. G., and Wong, F. L. (Eds.), *Geology and Offshore Resources of Pacific Island Arcs—Vanuatu Region*. Circum-Pac. Counc. Energy and Miner. Resour., Earth Sci. Ser., 8:287–300.

Johnson, D. P., Belford, D. J., Carter, A. N., and Crawford, A. J., 1988. Petrology and age of dredge samples collected in the Central Basin, Vanuatu. *In* Greene, H. G., and Wong, F. L. (Eds.), *Geology and Offshore Resources of Pacific Island Arcs—Vanuatu Region*. Circum-Pac. Counc. Energy and Miner. Resour., Earth Sci. Ser., 8:141–162.

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 Hebrides island arc. *Tech. Bull.—U. N. Econ. Soc. Comm. Asia Pac., Comm. Co-ord. Jt. Prospect Miner. Resour. South Pac. Offshore Areas*, 3:91–108.

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

Katz, H. R., 1988. Offshore geology of Vanuatu-previous work. *In* Greene, H. G., and Wong, F. L. (Eds.), *Geology and Offshore Resources of Pacific Island Arcs—Vanuatu Region*. Circum-Pac. Counc. Energy and Miner. Resour., Earth Sci. Ser., 8:93–124.

Kroenke, L. W., 1984. Introduction. *In* Kroenke, L. W. (Ed.), *Cenozoic Tectonic Development of the Southwest Pacific*. Tech. Bull.—U. N. Econ. Soc. Comm. Asia Pac., Comm. Co-ord. Jt. Prospect Miner. Resour. South Pac. Offshore Areas, 6:1–11.

Kroenke, L. W., Jouannic, C., and Woodward, P., (Compilers), 1983. Bathymetry of the Southwest Pacific, Chart 1. *Geophysical Atlas of the Southwest Pacific*. U. N. Econ. Soc. Comm. Asia Pac., CCOP/SOPAC Tech. Sec.

Louat, R., Hamburger, M., and Monzier, M., 1988. Shallow and intermediate-depth seismicity in the New Hebrides Arc: constraints on the subduction process. *In* Greene, H. G., and Wong, F. L. (Eds.), *Geology and Offshore Resources of Pacific Island Arcs—Vanuatu Region*. Circum-Pac. Counc. Energy and Miner. Resour., Earth Sci. Ser., 8:329–356.

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.

Macfarlane, A., Carney, J. N., Crawford, A. J., and Greene, H. G., 1988. Vanuatu—a review of the onshore geology. *In* Greene, H. G., and Wong, F. L. (Eds.), *Geology and Offshore Resources of Pacific Island Arcs—Vanuatu Region*. Circum-Pac. Counc. Energy and Miner. Resour., Earth Sci. Ser., 8:45–92.

Maillet, P. M., Monzier, M., Selo, M., and Storzer, D., 1983. The d'Entrecasteaux zone (southwest Pacific). A petrological and geochronological reappraisal. *Mar. Geol.*, 53:179–197.

Malahoff, A., Feden, R. H., and Fleming, H. S., 1982. Magnetic anomalies and tectonic fabric of marginal basins north of New Zealand. *J. Geophys. Res.*, 87:4109–4125.

Mallick, D.I.J., 1973. Some petrological and structural variations in the New Hebrides. *In* Coleman, P. J. (Ed.), *The Western Pacific: Island Arcs, Marginal Seas, Geochemistry:* New York (Crane, Russak and Co.), 193–211.

———, 1975. Development of the New Hebrides Archipelago. *Philos. Trans. R. Soc. London B*, 272:277–285.

Mallick, D.I.J., and Neef, G., 1974. Geology of Pentecost. *Regional Rept.—New Hebrides Geol. Surv.*

Mitchell, A.H.G., 1970. Facies of an early Miocene volcanic arc, Malakula Island, New Hebrides. *Sedimentology*, 14:201–243.

Mitchell, A.H.G., and Warden, A. J., 1971. Geological evolution of the New Hebrides island arc. *J. Geol. Soc. London*, 127:501–502.

Neef, G., Plimer, J. R., and Bottrill, R. S., 1985. Submarine-fan deposited sandstone and rudite in a mid-Cenozoic interarc basin in Maewo, Vanuatu (New Hebrides). *Sedimentology*, 32:519–542.

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.

Ravenne, C., Pascal, G., Dubois, J., Dugas, F., and Montadert, L., 1977. Model of young intra-oceanic arc: the New Hebrides island arc. *In Int. Symp. on Geodynamics in South-west Pacific, Noumea, New Caledonia, 1976*. Paris (Editions Technip), 63–78.

Shipboard Scientific Party, 1975. Site 286. *In* Andrews, J. E., Packham, G., et al., *Init. Repts. DSDP*, 30: Washington (U.S. Govt. Printing Office), 69–131.

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. [Correction in *J. Geophys. Res.*, 86:6066.]

Taylor, F. W., Jouannic, C., and Bloom, A. L., 1985. Quaternary uplift of the Torres Islands, northern New Hebrides frontal arc, comparison with Santo and Malakula Island, central New Hebrides frontal arc. *J. Geol.*, 93:419–438.

Wong, F. L., and Greene, H. G., 1988. Geologic hazards identified in the Central Basin region, Vanuatu. *In* Greene, H. G., and Wong, F. L. (Eds.), *Geology and Offshore Resources of Pacific Island Arcs—Vanuatu Region*. Circum-Pac. Counc. Energy and Miner. Resour., Earth Sci. Ser., 8:225–254.

Ms 134A-101