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# Environmental and tectonic influence on growth and internal structure of a fringing reef at Tasmaloum (SW Espiritu Santo, New Hebrides island arc, SW Pacific)

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## ABSTRACT

Subduction of the Australian Plate has caused rapid uplift of the central New Hebrides island arc (15°S, SW Pacific). The d'Entrecasteaux ridge system, a prominent bathymetric feature on the downgoing plate, is underthrusting the central part of the New Hebrides arc. The coastlines of most islands are characterized by emerged Holocene coral reef terraces. A maximum uplift rate of 6 mm yr$^{-1}$ occurs along the south-west coast of Espiritu Santo, near the plate boundary.

To investigate the Late Quaternary neotectonic and environmental evolution of the uplifted fringing reefs, we drilled the emerged Holocene reef at Tasmaloum (SW Espiritu Santo) to depths as great as 40–45 m. Coral samples from various levels were dated by $^{230}$Th and $^{14}$C, as described elsewhere, and the internal structure of the reefs was studied.

Preliminary palaeoecological and sedimentological data indicate the following. First, the coral reef colonized a substrate and began to grow by 24 ka on weakly indurated calcareous sand beds, which probably formed during the Late Pleistocene and are possibly as old as 30 ka or more. These sand levels could represent the deep fore-reef area of an older reef, at present behind the uplifted Holocene terraces. Second, the biofacies and coral ages from the reef sequence, which is continuous from the last glacial maximum (LGM), provide a view of the internal structure: between 24 and 12–10 ka, coral levels, composed mainly of *Acropora* gr. *hyacinthus* and gr. *cytherea* (accompanied by a few *Galaxea* gr. *fascicularis*), constitute medium- to high-energy assemblages, reflecting relatively deeper and more protected environments. However, at a few levels, acroporid build-ups of *Acropora* gr. *danai/robusta* indicate high-energy environment alternation. Between 12–10 ka and the present, assemblages of acroporids (*Acropora* gr. *danai/robusta*), scarce poritids, numerous encrusting coralline algae, vermetid gastropods and encrusting foraminiferids indicate high-energy environments, probably corresponding to the upper part of the exposed reef slopes. This biofacies succession indicates changes in wave energy, related to fluctuations in local bathymetry controlled by the net effect of variable rates of sea-level rise, reef growth and tectonic uplift. After 6 ka, the replacement of coral, coralline algal and stromatolite assemblages by coral and coralline algal frameworks implies the establishment of a new hydrological and oceanographic regime. The biofacies and age structure of the reef show that the uplift rate has varied since 24 ka.

## INTRODUCTION

Located in the tropical zone of the SW Pacific Ocean, the Vanuatu (New Hebrides) archipelago comprises the narrow volcanic island chain which stretches between 13°S (Torres Group in the north) and 20°S (Anatom in the South) and between 166°E and 172°E (Fig. 1). These islands are a part of the New Hebrides Island Arc (NHIA) at the convergent boundary of the Australian and Pacific Plates. This arc shows neotectonic features related to the subduction of ridges and seamounts. In particular, in its central part, there is no physiographical trench along the plate boundary west of Espiritu Santo, where the d'Entrecasteaux zone intersects the arc (Fig. 1), emerged islands in the forearc (Espiritu Santo and Malekula) occur anomalously close to the plate boundary and the backarc islands of Pentecost and Maewo have uplifted rapidly (Taylor *et al.*, 1987).

The maximum Late Quaternary uplift rates of Espiritu Santo (3–4 to 6 mm $yr^{-1}$ during the Holocene in its SW part) and North Malekula (maximum of 3.4 mm $yr^{-1}$ since 6 ka) coincide with the western topographical axis of these islands, which is subparallel to the trend of the trench (Jouannic *et al.*, 1980; Taylor *et al.*, 1980, 1985).

Previous studies of the coral reefs in Vanuatu

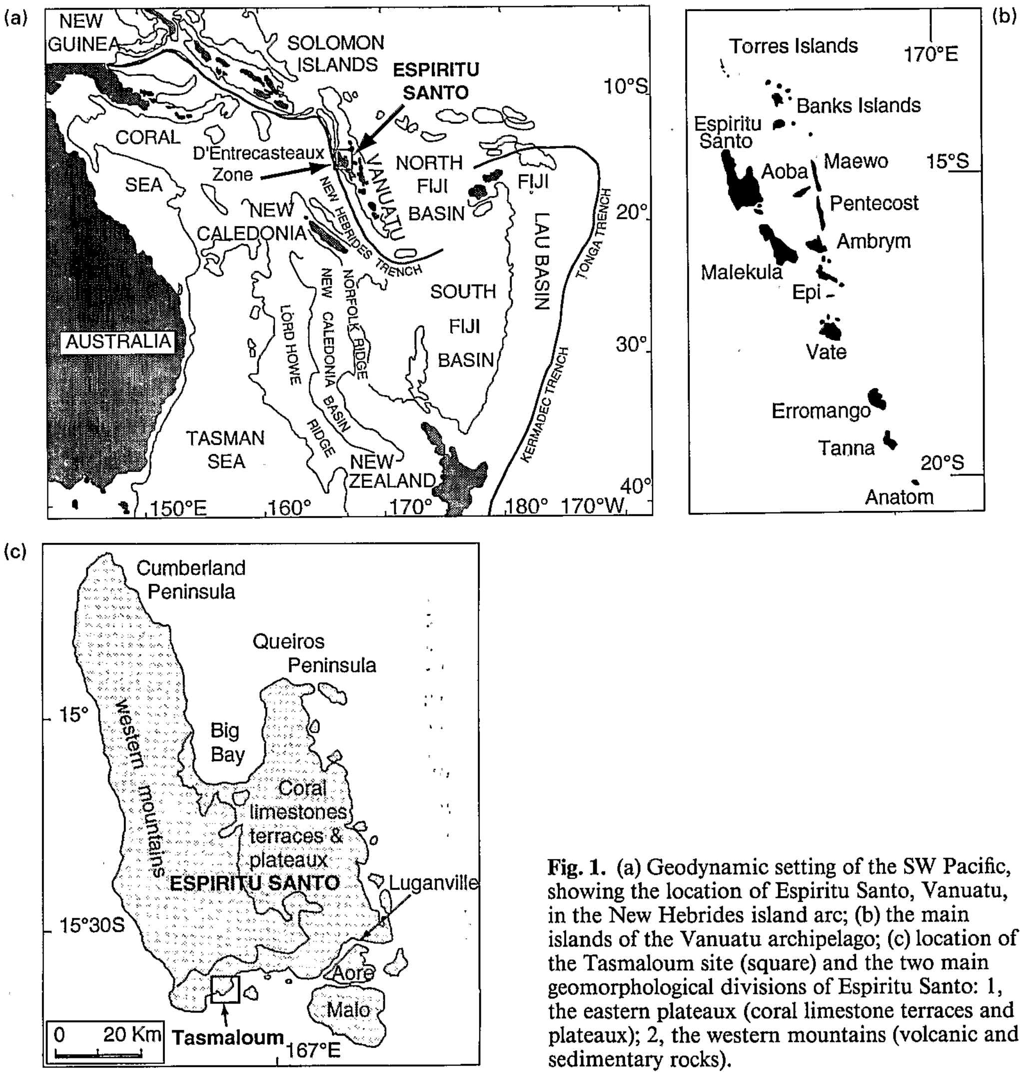

Fig. 1. (a) Geodynamic setting of the SW Pacific, showing the location of Espiritu Santo, Vanuatu, in the New Hebrides island arc; (b) the main islands of the Vanuatu archipelago; (c) location of the Tasmaloum site (square) and the two main geomorphological divisions of Espiritu Santo: 1, the eastern plateaux (coral limestone terraces and plateaux); 2, the western mountains (volcanic and sedimentary rocks).

have examined neotectonics, sedimentological features, palaeoenvironments and palaeoclimate. Guilcher (1974) studied the structure of reef types in Vanuatu, and identified fringing and open-sea reefs as primary types. Veron (1990) recognized about 296 species (belonging to 62 genera) of hermatypic corals in the archipelago. Done & Navin (1990) studied habitats of the shallow water communities of many coral reefs. Our studies have focused on neotectonics and palaeoclimate (Beck *et al.*, 1992; Gray *et al.*, 1993; Récy *et al.*, 1993; Taylor *et al.*, 1993).

In the Tasmaloum area (Fig. 1), a high uplift rate (6 mm $yr^{-1}$) was recorded along the coast at least for the last 6 ka (Taylor *et al.*, 1980; Gilpin, 1982; Bloom & Yonekura, 1985) and perhaps even higher for the last 12 ka (Taylor, 1992). Although this rate is high, we may assume that reef growth was continuous during a large part of Quaternary times, in the same way as on Huon Peninsula (Chappell & Polach, 1991; Edwards *et al.*, 1993). If so, we can obtain corals representing nearly all of the last 20 ka of reef growth by drilling relatively shallow cores. We calculated that drilling to depths less than 20–30 m would be sufficient. For example, if the uplift rate has been 6 mm $yr^{-1}$, then the 20-ka shoreline has been raised 120 m from its original level. Combining the uplift rate and total postglacial sea-level rise (e.g. Fairbanks, 1989), the net transgression at SW Espiritu Santo was hypothesized to have been only about 10 m. Based on this reasoning, we began a coring programme on the emerged part of the Tasmaloum fringing reef in 1990. We drilled several sites in 1990, 1992 and 1994 to sample a great number of pristine corals representing the past 20 ka.

Before our reef drilling programme, the growth history of Vanuatu fringing reefs was unknown. However, the internal structure of reef sites is documented in the SW Pacific, especially in Australia (Davies & Marshall, 1979; Hopley, 1982; Marshall & Davies, 1982; Collins *et al.*, 1993) and in New Caledonia (Cabioch *et al.*, 1995) and serves as a basis for comparison with the Vanuatu reefs.

The objective of this paper is to present the first palaeoecological and sedimentological results on the internal structure of a particular fringing reef (Tasmaloum, SW Espiritu Santo, Fig. 1), as well as preliminary geomorphological data concerning reef settlement and growth patterns. In addition, we interpret the data in the context of the environmental and tectonic features of the central NHIA.

## MATERIALS AND METHODS

Fourteen cores, obtained by vertical (9A to -G, -I, -M & -N) and oblique (9H, -J, -K & -L) drilling (Fig. 2), were recovered using a Jacro and, later, a Sedidrill coring system. The total depth to which we

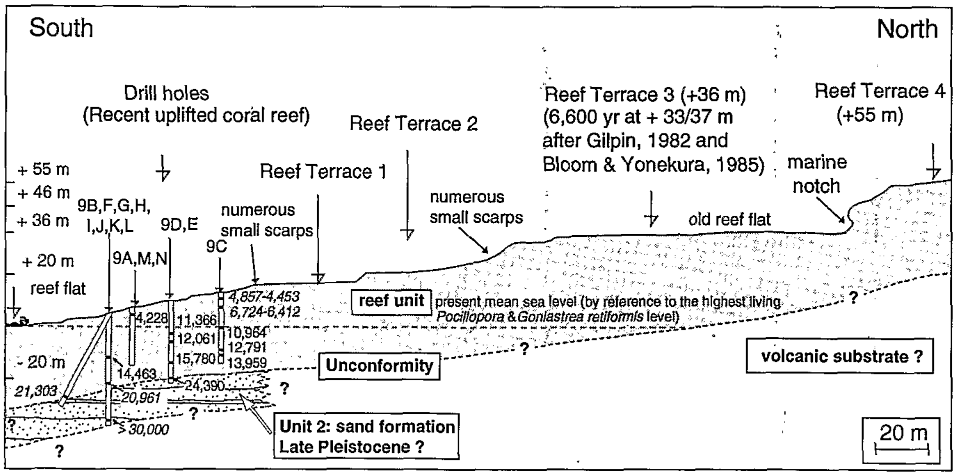

**Fig. 2.** Schematic cross-section of the uplifted Tasmaloum fringing reef, showing the main reef terraces and the two units recognized by drilling: (1) the Late Pleistocene (24 ka) to Holocene upper unit (coral debris and build-ups); (2) the Late Pleistocene (from, at least, 30 ka to 24 ka) lower unit (bioclastic sands). The deepest cores (vertical or inclined) are, at each drilling site, 9C, 9E, 9F, 9H and 9N. The inclined cores are 9H, 9J, 9K and 9L. The dates have been obtained through TIMS (U/Th dates) and AMS ($^{14}$C dates calibrated in calendar years, in italics).

drilled ranged from 15.25 to 41.35 m (Fig. 2), and the core diameters were 4.6–3.5 cm. Recovery ranged from 10 to 80%. Sections of poor recovery are assumed to be due to cavities or caverns, or may represent the occurrence of intervals of fine sand or unconsolidated sediments (Figs 2 and 3). During drilling, core sections were recovered every 1.50 m, thus permitting a depth accuracy of ±0.5 m. The depths (m) are given by reference (a indicates above; b indicates below) to the precisely levelled determinations of the highest living *Pocillopora* and *Goniastrea retiformis* level, which corresponds to the highest level of survival (HLS) of corals (e.g. Taylor *et al.*, 1987).

Core studies include geochemical, sedimentological and mineralogical analysis focusing on evaluating diagenesis, and biological aspects (fauna and flora components and associations). Samples were examined using standard petrographical binocular microscopy, scanning electron microscopy (SEM), X-ray diffraction and specific staining techniques. Semi-quantitative estimates were made from thin-sections to determine the composition of the biological assemblages associated with the coral framework.

Coral samples for isotopic dating were most commonly of *Porites* sp. or acroporids. We selected only corals which appeared pristine, containing more than 98% aragonite and showing minimal evidence of aragonite cement or dissolution. $^{230}$Th/$^{234}$U dates were obtained by thermal ionization mass spectrometry (TIMS) in the Minnesota laboratory, using the method developed by Edwards *et al.* (1987). The same coral samples were also dated by accelerator mass spectrometry (AMS) $^{14}$C (AMS facility, Tucson), to provide a calibration between the two methods (Gray *et al.*, 1993). Dates only obtained by $^{14}$C were converted to calendar years, using the calibration methods (Bard *et al.*, 1990, 1993; Edwards *et al.*, 1993; Stuiver & Reimer, 1993).

## REEF GEOMORPHOLOGY AT TASMALOUM

The coral reefs surrounding Espiritu Santo are generally fringing reefs. They are typically narrow, with reef fronts very close to the shores (Guilcher, 1974). They are not protected by barrier reefs or any other natural barrier, except near SE Espiritu Santo where small islands protect a few sites (Fig. 1). As a result, reefs are occasionally subjected to large waves, especially along the open SW coast where the Tasmaloum area is located (Fig. 1). The shore morphology immediately around the drilling area exhibits both narrow and broad terraces, in which gentle to steep slopes are observed (Fig. 2). On the uplifted reef, a succession of minor scarps are distinguishable.

Several broad terraces on the emerged Holocene reef have been formed successively between the present sea-level and the Holocene terrace level at about 36 m. The 36-m terrace level terminates against a scarp which rises to an upper terrace of older coral limestone at +55 m (Fig. 2). The +36-m reef terrace is the highest and broadest Holocene terrace, dated at 6.6 ka (Gilpin, 1982). A deep notch in the palaeo-sea cliff at the back of the 36-m terrace is observed. Except against this cliff, notches are generally missing from the scarps of the Holocene reef.

The present uplifting reef flat near the drilling sites shows typical reef zonation: an outer reef zone occupied by small coral colonies (generally acroporid forms) with encrustations of coralline algae; a median reef zone with many branching corals and somewhat indurated fine bioclastic sediments; and a back-reef zone having few large coral colonies and weakly indurated bioclastic sediments. Behind, the supratidal zone consists of an ancient and very gently sloping uplifted reef surface. This zonation is similar to that typically encountered in the south-west Pacific, i.e. in modern Vanuatu coral reefs (Done & Navin, 1990), in many neighbouring New Caledonian (Cabioch *et al.*, 1995) and Solomon (Morton & Challis, 1969) fringing reefs, and in the Australian Great Barrier Reef (Partain & Hopley, 1989).

The modern fringing reef is narrow, and living corals are restricted to the reef front, the seaward part of the reef flat and the reef slope. Robust, wave-resistant organisms, well adapted to strong water movement, and coralline algal veneers are frequently present in this high-energy environment. Bioeroders are common and constitute good markers of sea-level, especially the *Echinometra* level near mean sea-level. The modern fore-reef slope is steep.

## TASMALOUM INTERNAL REEF STRUCTURE

Cores drilled at Tasmaloum penetrated the Holocene, Late Pleistocene and LGM reef sequence back to 24 ka (Fig. 2), and recovered the underlying reef substrate. Coral reefs that grew during the postglacial transgression colonized a substrate of calcareous sands, gravel and pebbles (Fig. 2).

## Reef substratum

The substrate upon which the Tasmaloum fringing reef is established consists of sand layers, found only in the deepest cores, i.e. 9E, 9F, 9G and 9H (Fig. 2). This sand substrate (older than 24 ka and probably as old as, at least, 30 ka) consists of fragments, mainly of molluscs (bivalves and gastropods), *Halimeda* and benthic Foraminifera, represented especially by *Amphistegina* spp., accompanied by *A. lessoni* and *A. radiata* (the most abundant), *Alveolinella quoyii* or *Borelis schlumbergeri*, *Calcarina* sp. at −35 m bHLS and −37.50 m bHLS. Green algae, *Halimeda* sp., are particularly abundant at around −36.50 m bHLS.

## Reef biofacies sequences

Corals recovered in the cores are dominated by acroporids and poritids (Figs 2 & 3). Coralline algae belong usually to *Porolithon* cf. *onkodes*, *Porolithon* sp., *Neogoniolithon* spp., *Neogoniolithon fosliei*, *Dermatolithon tesselatum*, *Lithophyllum* sp.,

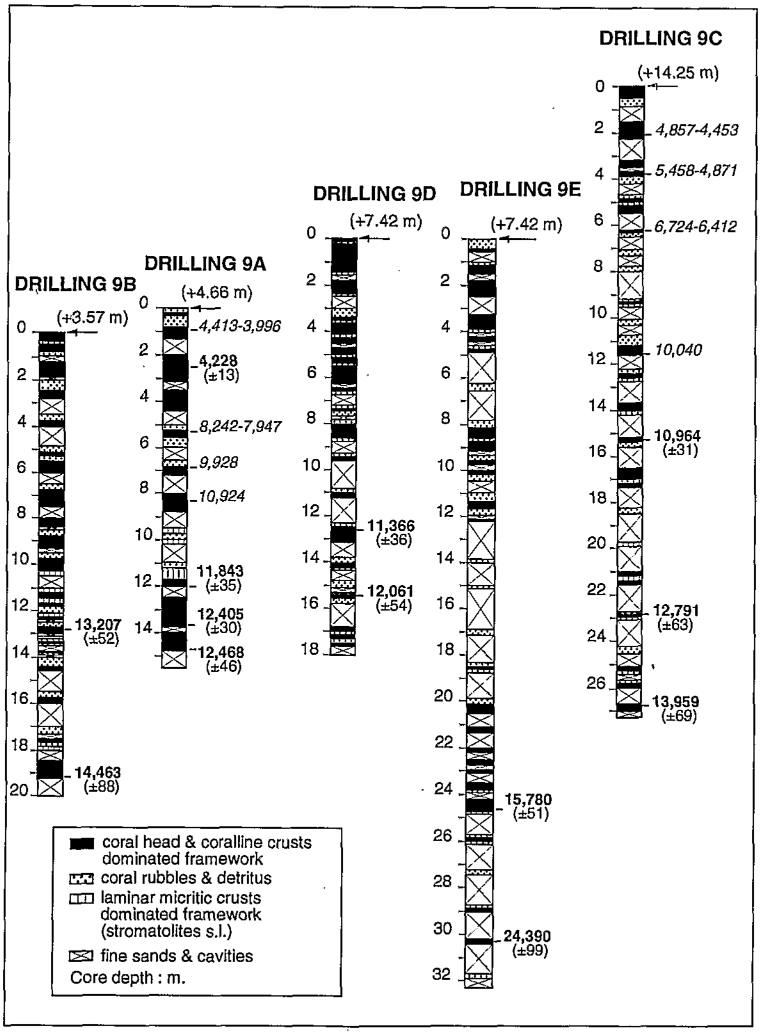

**Fig. 3.** Lithological features of the most characteristic vertical cores. Italic type indicates the $^{14}$C dates, calibrated in calendar years, and bold type indicates the TIMS U/Th dates (2σ precision in parentheses). The top of each drilling is given by reference to HLS.

*Tenarea* sp., and various other forms. The main coral-facies of the 24 ka Tasmaloum reef sequence are distributed within two distinct units (Fig. 4):

1 A lower unit, from 24 to 12–10 ka (between -26 and -5 m bHLS), is composed of branching coral facies with *Acropora* sp. gr. *hyacinthus*, *Acropora* sp. gr. *cytherea*, various other acroporids, and of coral heads with faviids such as *Montipora digitata* and sometimes *Galaxea* gr. *fascicularis*. Nevertheless, in several cores, *Acropora* sp. gr. *danai* and various acroporids are found. In this lower unit, the coralline algae (generally *Porolithon onkodes* and/or *Dermatolithon* cf. *tesselatum*) are scarce. From 15 to 12–10 ka (between -14 and -5 m bHLS), acroporids are abundant, composed usually of *Acropora* gr. *danai* and branches of *Acropora* sp., accompanied by scarce faviids.

2 An upper unit, from 12–10 ka to the present, consists of *in situ* massive *Porites* facies and a few faviids, alternating with beds of *Acropora* fragments (i.e. coral rubble facies) or *Acropora* sp. gr. *danai*, or various acroporids. Several coralline algal levels are composed of *Porolithon* cf. *onkodes* and/or *Neogoniolithon* spp. (or *Neogoniolithon fosliei*), and/or

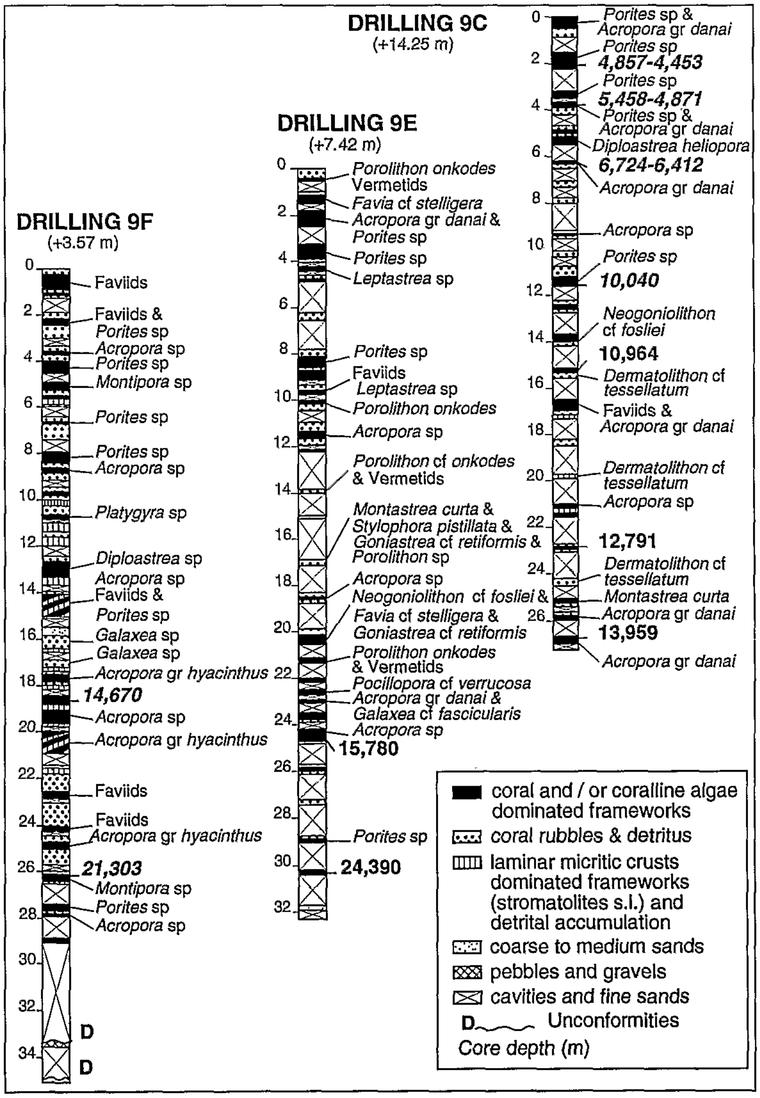

**Fig. 4.** Main corals (faviid, poritid and acroporid forms), coralline algae (species of *Porolithon*, *Neogoniolithon*, *Dermatolithon*) and vermetid gastropods recognized in cores 9B, 9C, 9E and 9F. The dates in bold type have been obtained through TIMS and those in bold italic type through AMS calibrated in calendar years. The top of each drilling is given by reference to HLS.

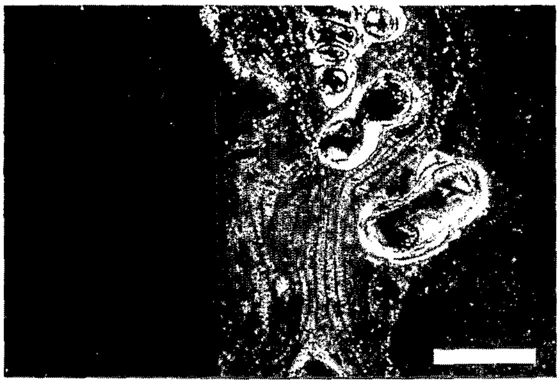

**Fig. 5.** Thin-section photomicrograph of *Dermatolithon* cf. *tesselatum*, *Porolithon onkodes* (coralline algae) and vermetid assemblage at −15.50-m core depth (i.e. 11.93 m bHLS) in core 9B, typical of a high-energy environment (scale bar represents 500 μm).

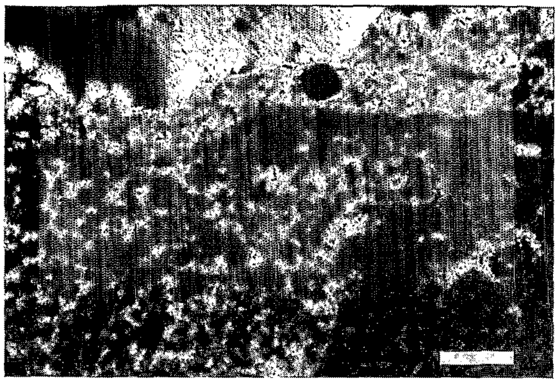

**Fig. 6.** Thin-section photomicrograph of a peloidal high-magnesian calcite cement partially infilling a cavity in core 9B at −18.60-m core depth (i.e. 15.03 m bHLS). Peloids are rimmed by a high-magnesian calcite fibrous to bladed cement. The clotted texture of the peloidal cement is less dense at the top than at the bottom of the cavity (scale bar represents 160 μm).

*Dermatolithon tesselatum* in which vermetid gastropod borings (generally *Dendropoma*) are usually intermingled (Fig. 5). Several layers of encrusting coralline algae are commonly observed around coral debris.

The detritus, generally uncommon except for the cavity infillings, consists of corals, green algal particles (*Halimeda* sp.), benthic foraminiferids, red algae and molluscs. This biofacies usually occurs near the top of the cores.

## Lithification

### *Micritic infilling and cements*

Thin-section studies reveal numerous borings infilled by micrite and internal sediments, particularly abundant in the upper metres of the cores. The bioclasts are usually cemented by high-magnesium calcite, formed generally of peloidal micrite (Fig. 6), a type described in reefs by Macintyre (1977) and Marshall (1986). Coral framework, coralline algal veneers and most bioclasts (molluscs, etc.) are bored by various organisms, especially fungi, sponges (clionids) and algae. In most cases, these borings and most other cavities (in particular, the primary pores) are infilled by micritic sediments, admixed with very fine debris of corals, algae, molluscs, and sometimes by geopetal pelleted high-magnesian calcite. Isopachous high-magnesium fibrous cement occurs occasionally, indicating cements of marine origin (Macintyre, 1977; Aissaoui & Purser, 1986; Marshall, 1986; Aïssaoui, 1988). In a few cases, botryoidal aragonite cements link the coralline algae layers.

A particular sedimentological feature is the partial micritization of mollusc and coralline algal fragments, especially in the upper metres of the cores. The complete replacement of the fragments by micrite is usually not observed, except in some red algal layers. These micritizations are probably due to abundant borings observed in these pieces (Bathurst, 1966). Although the degree of lithification depends on the skeletal framework (Aissaoui & Purser, 1986; Macintyre & Marshall, 1988), a great extent of submarine cementation is observed from the bottom to the top of cores, independent of the reef biofacies. This is probably due to the external position of the reef build-ups in a high-energy water environment, as reported elsewhere. A few dark minerals, infilling cavities, are widely observed throughout the cores, indicating inputs of terrestrial sediments from the island.

### *Laminar micritic crusts*

In most of the cores, from around 5–6 ka (at +3 m aHLS) to around 20 ka (at −23 m bHLS), high-magnesian micritic calcite laminar crusts, clotted or peloidal in aspect, are observed (Fig. 7). These laminar micritic crusts are especially abundant from −6.50 m bHLS (at 11–10 ka) to −18.50 m

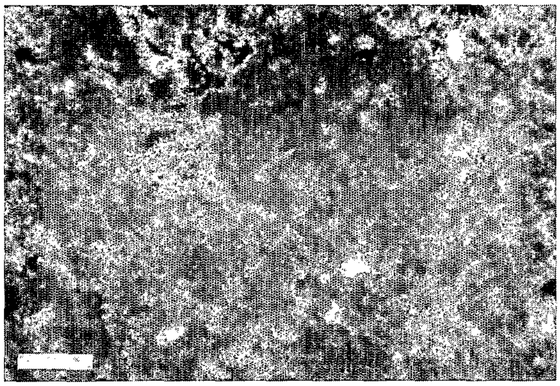

**Fig. 7.** Thin-section photomicrograph of a micritic laminar crust at -11.80-m core depth (i.e. 8.23 m bHLS) in core 9B. The voids and micritic reticulate network are typical of an organic edification (scale bar represents 160 μm).

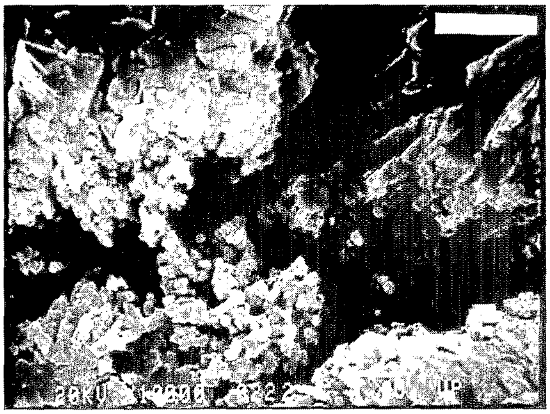

**Fig. 9.** SEM view of a micritic plan-laminar crust at -10-m core depth (i.e. 5.34 m bHLS) in core 9A, showing high-magnesian calcite cement and grape-shaped forms attributed to microbial remains (scale bar represents 2 μm).

bHLS (at 16 ka) (Figs 3 and 4), and seem to disappear after 6 ka. The laminar crusts are usually smooth and generally occur intermixed with corals and coralline algae, forming a typical biological succession. Corals in growth position, usually *Acropora* gr. *danai/robusta* (or various acroporids) indicating high-energy conditions, are coated by coralline algae. In most cases, these encrusting red algae are associated with vermetid gastropods (Fig. 5), and seldom with encrusting foraminiferids. Above these encrustations, micritic plan-laminar coatings are found, usually up to 5 cm thick. These laminar micritic coatings consist of peloidal and dense micrite (Fig. 7), forming wavy laminae and, sometimes, a dense micritic network is observed (Fig. 8). Occasionally, many well-defined stratified levels of bioclastic fragments are interbedded in these laminae. Grape-shaped aggregates, similar to bacterial remains (Figs 9 & 10), are attributed to organic constructions similar to the stromatolites described in Polynesian high-energy barrier reef

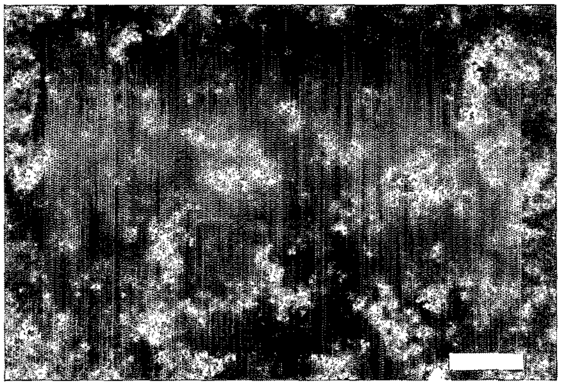

**Fig. 8.** Thin-section photomicrograph of dense micritic tubes in a plan-laminar micritic crust at -5.40-m core depth (i.e. 0.74 m bHLS) in core 9A. In this dense micritic crust, a polygonal form is observed. This form could represent filament remains characteristic of an organic activity (scale bar represents 80 μm).

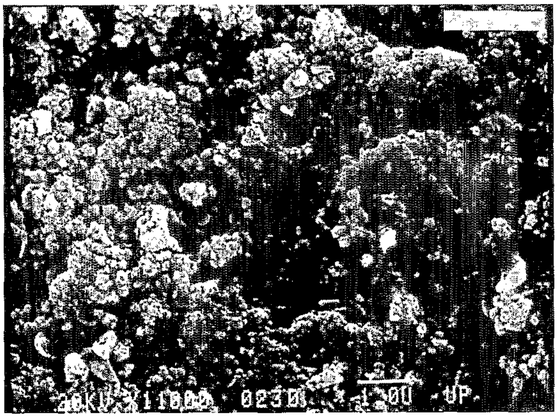

**Fig. 10.** SEM view of a micritic plan-laminar crust in which many grape-shaped forms are recognized, similar to bodies of microbial remains (at -17.20-m core depth, i.e. 2.95 m bHLS, in core 9C) (scale bar represents 2 μm).

**Table 1.** $\delta^{18}O$ and $\delta^{13}C$ values of laminated stromatolitic crusts of cores 9A, 9B and 9C at various depths

| Core number | Sample number | Depth* (m) | $\delta^{18}O$ (‰ PDB) | $\delta^{13}C$ (‰ PDB) |
|---|---|---|---|---|
| 9A | 202 | -4.85 | 0.05 | 3.77 |
| | 201 | -5.35 | -0.02 | 3.78 |
| | va 32 | -6.85 | 0.08 | 3.67 |
| 9B | 129 | -7.50 | 0.28 | 3.61 |
| | 165 | -10.20 | 0.23 | 3.66 |
| | 206 | -14.30 | 0.12 | 3.11 |
| 9C | 143 | -3.0 | -0.13 | 3.44 |
| | 151 | -5.55 | 0.11 | 3.41 |
| | 176 | -10.95 | -0.71 | 2.15 |

*Depth below the highest living *Pocillopora* and *Goniastrea retiformis* level.

environments (Montaggioni & Camoin, 1993; Camoin & Montaggioni, 1994). Values of $\delta^{18}O$, from -0.02 to 0.28, and $\delta^{13}C$, from 2.15 to 3.78 (Table 1), are typical of organic fractionation and confirm this assumption. The Vanuatu crust values are close to the values found in the Polynesian stromatolitic crusts (Fig. 11). Nevertheless, a shift toward positive values in $\delta^{18}O$ is observed. In summary, morphology, structure and isotopic data of the laminar micritic crusts indicate an organic origin, as pointed out in Tahiti (Camoin & Montaggioni, 1994).

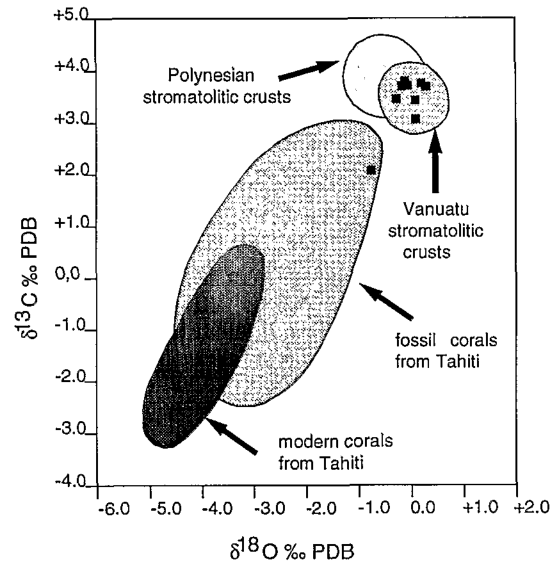

**Fig. 11.** Isotopic data values of $\delta^{18}O$ measured in the laminar micritic veneers (■, see values in Table 1). For comparison, the $\delta^{18}O$ and $\delta^{13}C$ isotopic values of Polynesian stromatolitic crusts and Polynesian modern and fossil *Acropora* branches are taken from Camoin & Montaggioni (1994).

### Freshwater cementation and diagenetic changes in bioclastic fragments

In a few cases, oxidized, grey–brown silt-size infillings and partial dissolution of aragonitic components are observed from the bottom to the top of the cores, indicating emergence events and meteoric water throughflow. Also, matrix dissolution occurs around the bioclasts, especially at the top of the cores (Fig. 12). Coral heads or coral branches are not much affected by freshwater leaching. Sometimes, although corals seem to be pristine (conservation of aragonitic fabrics), SEM observations reveal solution features (Fig. 13) in which the leaching of sclerodermite centres is observed. Dissolution of

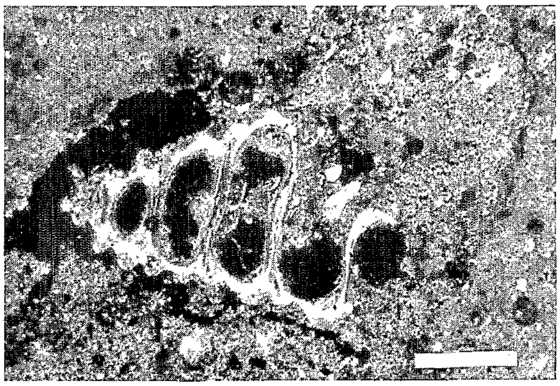

**Fig. 12.** Thin-section photomicrograph of micritic matrix dissolution around a gastropod at -0.80-m core depth (i.e. 2.72 m aHLS), in core 9B (scale bar represents 500 µm).

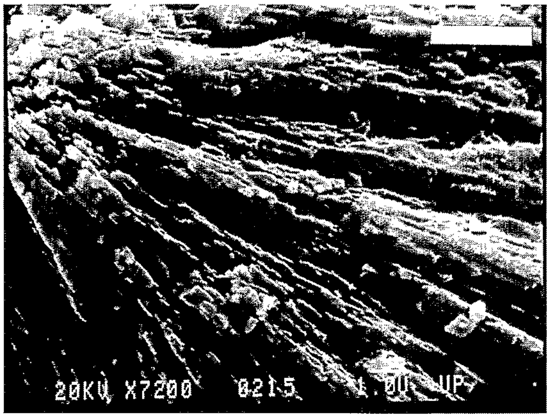

**Fig. 13.** SEM view of a *Porites* sample from the upper metres of core 9A, showing partial leaching of the aragonitic needles. The centre of the sclerodermite is clearly affected by dissolution (scale bar represents 3 μm).

**Fig. 14.** SEM view of a *Porites* sample (see Fig. 8) from the upper metres of core 9A, in which we can recognize some low-magnesian calcite crystals (euhedral form), indicating freshwater diagenesis: the aragonitic sclerodermite leaching is followed by a partial low-magnesian calcite cement replacement (scale bar represents 4 μm).

aragonite needles is also found in the corals (Fig. 13), at both the bottom and top of cores. Most rarely, low-magnesian cement fills voids in coral (Fig. 14). Such a leaching sequence is interpreted as the first stage of progressive freshwater diagenesis (James, 1974; Gvirtzman & Friedman, 1977; Henrich & Wefer, 1986).

## DISCUSSION

### Morphology

The narrow Holocene terraces along the Tasmaloum peninsula, separated by gentle slopes without notches (except at the back of the 36-m terrace), may indicate continual rapid uplift movements with only brief intervening pauses. Likewise, the pronounced steepness of the slope that is observed at present along the Tasmaloum shore indicates a continued process of frequent successive uplift movements resulting in a rapid mean uplift rate. Several researchers have assumed that the small terraces and scarps could represent coseismic uplift events (Ota *et al.*, 1993; Chappell *et al.*, 1994; Ota & Chappell, 1996). However, the observations previously made in Vanuatu after the 1965 uplift of North Malakula, which raised corals by as much as 1.2 m (Benoit & Dubois, 1971; Taylor *et al.*, 1980, 1987), did not show any kind of distinctive terrace or scarp building. The same observation has been made in the Tasmaloum area, where the same 1965 seismic event uplifted the coast by about 0.25 m (Taylor *et al.*, 1980, 1987). Building of small terraces and scarps could be due to simultaneous combination of coral reef growth especially when the reefs arrive in the intertidal zone because of uplift, bioerosion and solution processes, sea-level rise and succession of more or less numerous tectonic uplift events, such as those previously reported in other parts of the archipelago (Benoit & Dubois, 1971; Jouannic *et al.*, 1980; Taylor *et al.*, 1980, 1987, 1990), or in Papua New Guinea (Bloom *et al.*, 1974; Chappell, 1974; Chappell & Polach, 1991).

Formation of the deep notch at the back of the 36-m terrace, dated at *c.* 6 ka, is directly related to growth of the broad 36-m reef flat and the relative stability of sea-level at a time when the rates of uplift and rising sea-level equalized for some time (Fig. 2). Absence of other representative notches from the scarps of the Holocene reef below the 36-m terrace (Fig. 2) indicates that, since 6 ka, the position of the shoreline has regressed nearly continuously, so that dissolution processes and endolithic organism borings were never at one level long enough to form other notches (Pirazzoli, 1986).

## Pre-reef substrate, reef colonization and initial growth

Colonization of the Tasmaloum fringing reef occurred upon a pre-reef substrate, encountered in the cores at depths of -20 to -25 m bHLS (Figs 2 & 15). This substrate is composed of a huge bioclastic sand formation accumulated during a part of the last glacial period. The faunistic (especially the benthic foraminiferids) and floristic assemblages of these sand levels are typical of an algal environment. The apparent absence of corals (colonies or debris) seems to be in accordance with the fact that the sand beds could represent the fore-reef of an older reef just behind the present Holocene terraces. A combination of sea-level fall or rise and tectonic uplift eventually produced a substrate with a depth suitable for colonization and reef development. However, the substrate became shallow so fast that coral growth was apparently prevented; this is consistent with the lack of coral before 24 ka.

Above the basal sand formation, pebbles and gravel conglomerates indicate a palaeo-coastline (Fig. 15). The reef growth initiation was almost instantaneous, as shown by similar dates obtained in core 9F at -22.65 m bHLS (20 961 cal yr BP) and in core 9J at -28.25 m bHLS (21 303 cal yr BP). Nevertheless, in core 9E, a date of 24 390 cal yr BP was obtained at -22.70 m bHLS. The earliest stages of Tasmaloum reef initiation seem to have been contemporaneous with the late stages of the LGM and the earliest stages of the sea-level rise, at around 20 ka. Previous to this study, Holocene reef initiations were observed to occur in the Pacific at around 9–7 ka (Easton & Olson, 1976; Davies & Montaggioni, 1985; Davies *et al.*, 1985, Montaggioni, 1988; Cabioch *et al.*, 1995; Kan *et al.*, 1995; Larcombe *et al.*, 1995), and in the barrier reef in Tahiti at around 14 ka (Bard *et al.*, 1996). Variations in reef initiation ages of these various areas could be explained by sea surface temperature (SST) differences unfavourable to reef development or the unavailability of suitable foundations for reef establishment (Davies *et al.*, 1985; Partain & Hopley, 1989). Near Tasmaloum, located at 15°S, at around 24 ka, SSTs were probably high enough for the reef growth. Also, the gravels and pebbles provided a favourable substrate for reef initiation, as previously observed in Australia (Johnson & Risk, 1987; Partain & Hopley, 1989). In New Caledonia, we know that the time lag for Holocene coral colonization between reef areas is related to differences of substrate type, fringing reefs colonizing preferentially karstified limestone foundations (or other similar surfaces), the roughness of karstic surfaces (or similar) being favourable to recruitment and attachment of coral larvae (Cabioch *et al.*, 1995).

At Tasmaloum, the morphology of the underlying pre-reefal substrate (top of the sand accumulations, Figs 2 & 15) influenced the earliest stages of reef

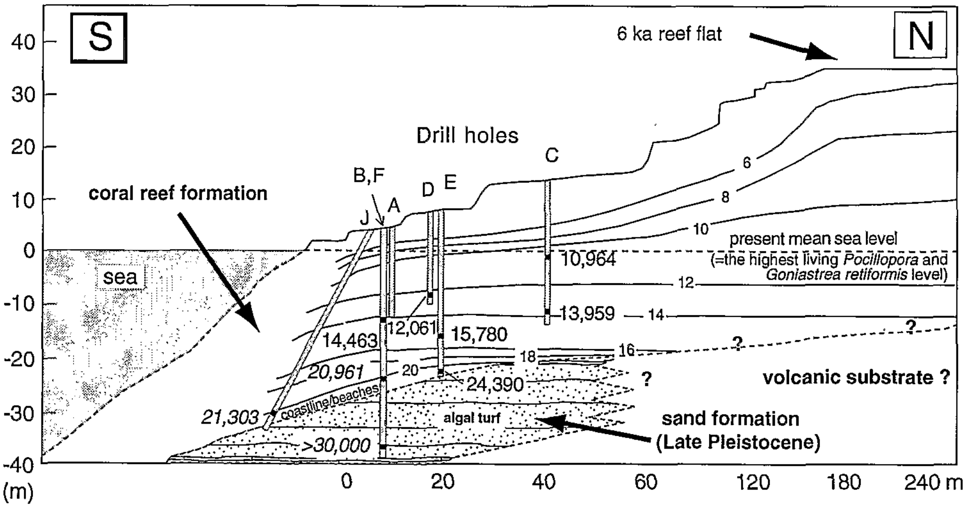

**Fig. 15.** Cross-section of the uplifted Tasmaloum fringing reef, showing the time-lines of 6, 8, 10, 12, 14, 16, 18 and 20 ka. Only the most characteristic dates are reported (those in italics are $^{14}C$ dates converted to calendar years).

development, as indicated by the regularity of the time-lines between 24 ka and 12–10 ka (Fig. 15), which is also usually observed in the younger reefs (Marshall & Davies, 1982; Walbran, 1994). After 12–10 ka, the tectonic behaviour of Espiritu Santo constrained the modern structure. This gives an explanation for the relatively narrow reef and steep reef slope (Bloom & Yonekura, 1985), similar to the situation in Papua New Guinea (Chappell, 1974; Bloom *et al.*, 1974).

## Palaeoecological significance

The environmental factors which influence reef growth are directly related to water depth, light intensity, wave energy, nutrients, sediment and freshwater inputs: characteristic assemblages of organisms represent well-defined zones, as pointed out elsewhere in the south-west Pacific (Morton, 1973, 1974; Guilcher, 1974; Done & Navin, 1990). Consequently, coral and organism assemblages found in the cores commonly have a well-defined palaeoecological significance (biozonal partitioning) and we can use these to reconstitute the various stages of reef development.

The coral assemblage of the lower unit (from 24 to 12–10 ka) is principally composed of branching coral facies and various acroporids, *Acropora* spp., as generally observed in the Pacific (Davies *et al.*, 1985; Cabioch, 1988; Montaggioni, 1988; Veron, 1990). The most frequent *Acropora*, *A.* gr. *hyacinthus* and *A.* gr. *cytherea*, are common in environments having heavy waves and surges and in the central part of reef slopes (Morton, 1974; Faure, 1982; Veron, 1990). This coral assemblage, accompanied by scarce coralline algae, indicates medium- to high-energy conditions developed predominantly in relatively deep or protected environments (Done, 1982; Faure, 1982; Veron, 1990), perhaps related to the 5–15-m reef slopes. The textural analysis confirms this interpretation. Furthermore, from 15 to 12 ka, a *A.* gr. *danai/robusta* and *Porolithon* cf. *onkodes* community, sometimes found mixed with the previous acroporid assemblage, is typical of high wave energy (Faure, 1982; Veron, 1990). The second most frequent forms, poritids, are abundant, but are found everywhere, and thus are not good bathymetric indicators. Faviids are common, but never abundant, except in core 9B, where a large colony of *Diploastrea heliopora* is observed.

The upper unit (between 12–10 ka and the present) is composed principally of *Acropora* sp. gr. *danai*, or various other strong branching acroporids, typical of high-energy conditions, probably related to the reef front and the upper reef slope (Done, 1982; Faure, 1982; Veron, 1990). *In situ* massive *Porites* sp. are more common. We note that in core 9D, at around −1 m HLS, the coral assemblage is made of *Favites* cf. *pentagona*, *Seriatopora* sp. and *Pachyseris rugosa*, which indicates a moderate-energy coral reef environment.

In these cores, the occurrence at many levels of heavy coralline red algal crusts, especially in the upper unit (generally *Porolithon* cf. *onkodes*), is very closely linked to local environmental conditions, especially wave energy and light intensity (Littler & Doty, 1975; Bosence, 1984, 1985). In these thick coralline algal crusts, vermetid gastropods (generally *Dendropoma maxima*) and encrusting foraminiferids are common (Fig. 4). In the Pacific, this specific assemblage is typical of high-wave-energy environments on hard substrates, generally at depths less than 2–6 m below mean sea-level (i.e. near the reef front or at the top of the upper reef slopes). In particular, *Porolithon onkodes* requires intense light and continuous disturbance (Morton, 1973, 1974; Littler & Doty, 1975; Richard, 1982; Adey, 1986; Nunn, 1993). Sometimes, *Porolithon* cf. *onkodes* is found to be associated with *Neogoniolithon fosliei* or *Dermatolithon* cf. *tesselatum* in the uppermost metres of the cores (Fig. 4), these species being sharply restricted to depths less than 10 m (Adey *et al.*, 1982).

## Biofacies succession, reef development and neotectonic behaviour

During the postglacial sea-level rise, the vertical succession of reef assemblages generally reflects environment and bathymetry variations. These changes are related to antecedent topography and variations in wave energy (Davies & Hopley, 1983; Montaggioni, 1988; Collins *et al.*, 1993). Similar vertical variations occurred in the Tasmaloum fringing reefs during the last 24 kyr.

From 24 to 15 ka, the water-energy conditions suggest that reef growth at Tasmaloum did not keep up with sea-level rise. However, from 15 to 12–10 ka, the intermingled high-wave-energy assemblage (*A.* gr. *danai/robusta* community) seems to indicate an alternation between catch-up and keep-up biofacies in the sense of Neumann & Macintyre (1985). These community variations are probably related to uplift events.

From 12–10 ka to the present, the high-wave-energy assemblages demonstrate that the reef remained close to sea-level. Thus, the Tasmaloum fringing reef can be regarded as a keep-up reef between 12–10 ka and the present, because reef growth was able to keep pace with rising sea-level, or, alternatively, because successive uplifts permitted it to keep pace with sea-level rise, ranging from 0- to 6-m water depth. At about 5–6 ka, the sea-level stabilized, as reported elsewhere in the Pacific (Lambeck & Nakiboglu, 1986), but the uplift continued (Jouannic *et al.*, 1980; Taylor *et al.*, 1980, 1987) and thus terraces, scarps and notches formed (Fig. 2).

The bathymetry variation, observed in the internal structure of the Tasmaloum fringing reef, shows also that the uplift rate may not have been constant during the last 24 ka. Changes in reef growth patterns, particularly well characterized from 15 to 12–10 ka, imply an increase in uplift rate. In SW Espiritu Santo, this increase in uplift rate is probably related to the collision of the d'Entrecasteaux Zone with the New Hebrides arc.

## Lithification

### *Physico-chemical lithification and freshwater diagenesis*

The Tasmaloum fringing reef is well lithified by silt-size sediment infillings and, more importantly, by pelleted micritic high-magnesian calcite, especially in the uppermost metres of the cores. As reported by numerous workers, these well-lithified materials are restricted to outer reef environments, exposed to heavy wave action (Aissaoui & Purser, 1986; Cabioch, 1988; Macintyre & Marshall, 1988). This agrees with the implications of biological assemblages typical of medium- to high-energy water conditions, which represent the reef front (upper unit) or the reef slopes (upper and lower units).

Sometimes, freshwater diagenesis, revealed by discrete dissolution features, low-magnesian calcite cements (Fig. 14) and rare dissolution moulds, clearly shows that reef growth was disturbed by short periods of emergence. These are accompanied by freshwater inputs from episodic and successive uplift movements or pauses in sea-level rise related to deglaciation. Although leaching is observed, these reef materials are usually consolidated by freshwater cementation with low-magnesian calcite spar cement or micrite infillings.

### Organic lithification and stromatolites

In the Tasmaloum fringing reef, organic lithification is one of the most interesting sedimentological features, and numerous stromatolitic coatings are observed. The Tasmaloum reef is exposed to frequent heavy surf, similar to the Polynesian high-energy barrier reef (Montaggioni & Camoin, 1993). In neighbouring New Caledonia, such micritic coatings are rare or absent. Nevertheless, we observe similar thin laminations in the Holocene fabric of a lagoonal reef located near the barrier reef, where high-wave-energy conditions prevail. These data suggest that high-wave-energy conditions are required for reef stromatolitic development.

In the Tasmaloum reef cores, the stromatolites are found in abundance from 16–15 to 12–10 ka, a period of very rapid sea-level rise (Fairbanks, 1989; Bard *et al.*, 1990, 1996; Edwards *et al.*, 1993), in which palaeoceanographic parameters (SSTs and perhaps nutrient inputs) varied significantly. Another interesting fact is the disappearance of these laminar crusts after 6 ka. This corresponds to the time of sea-level stabilization (Lambeck & Nakiboglu, 1986), which constituted a generally favourable period for reef growth (Davies *et al.*, 1985; Montaggioni, 1988). Sea-level stabilization caused the establishment of a new hydrological and oceanographic regime, accompanied by, in particular, warming of tropical waters (Beck *et al.*, 1992, 1997). After 6 ka, conditions became more favourable for coral growth, but unfavourable for stromatolite growth. This is particularly well explained by changes in the SSTs recorded by isotopic methods in many *Porites* from Tasmaloum (Beck *et al.*, 1992; Récy *et al.*, 1993; Beck *et al.*, 1997). The Sr/Ca palaeo-SSTs, measured by TIMS, are very close to present-day values in the youngest corals (< 5 ka), but several degrees lower than present-day values in the other corals (between 5 and 10.3 ka). These recent results suggest also that the tropical belt was compressed toward the Equator before 10 ka (Beck *et al.*, 1992). Another explanation of these ecological changes may be related to palaeoceanographic changes, such as those pointed out recently by McCulloch *et al.* (1996).

## CONCLUSIONS

At Tasmaloum, morphological, palaeoecological and sedimentological data show the initiation of a fringing reef along an uplifted coast and an example of unusual reef growth on a rapidly uplifting substrate.

**1** Although SSTs before 10 ka were probably lower than at present, as inferred by Beck *et al.* (1992), and coral build-ups less numerous than they are today, the reef colonized a more-or-less indurated substrate (calcareous sands) by 24 ka (Gray *et al.*, 1993) and continued to grow until now.

**2** We have cored a reef sequence as complete as the Barbados section (Fairbanks, 1989; Bard *et al.*, 1990) and more complete than the Huon Peninsula (only the last 13 ka were recovered in that case; Chappell & Polach, 1991; Edwards *et al.*, 1993) and French Polynesian (only the last 14 ka; Bard *et al.*, 1996) sections.

**3** The biofacies analysis reveals two distinct units:

(a) From 24 to 12–10 ka, rare coralline algae and poritid and acroporid build-ups of *Acropora* gr. *hyacinthus* and *A.* gr. *cytherea* are considered to represent a medium- to high-wave-energy facies, related to 5–15-m water depth. Nevertheless, from 15 to 12–10 ka, this biofacies alternates with a high-wave-energy facies, composed mainly of *Acropora* gr. *danai/robusta*, implying several uplift events.

(b) From 12–10 ka to the present, assemblages of acroporids (*Acropora* gr. *danai/robusta*), rare poritids, numerous encrusting coralline algae (*Porolithon* cf. *onkodes*, *Neogoniolithon* spp., *Neogoniolithon fosliei* and *Dermatolithon tesselatum*), vermetid gastropods, and sometimes *Acervulina* foraminiferids reveal a high-wave-energy facies, corresponding to the upper part of the exposed reef slope or the outer part of the fringing reef flat, i.e. ranging from 0- to 6-m water depth. From 6–5 ka to the present, reef growth is marked by a succession of emergences resulting from the combination of sea-level stabilization and incremental uplift movements.

**4** Patterns of Tasmaloum reef growth appear to depend on the tectonic behaviour of Espiritu Santo in relation to the uplift movements and the more-or-less rapid rises or stillstands of postglacial sea-level. Thus, it is inferred that the uplift rate has varied during the last 24 ka (from 24 ka to 4.5 ka). After 6 ka, the vertical biofacies change, marked by the disappearance of the laminar crusts (or stromatolites sensu lato), reflects the establishment of a new hydrological and oceanographic regime.

**5** Once sea-level stabilized, the wide 6–7-ka reef flat at +36 m quickly emerged by uplift. The reef then had only the steep transgressive reef-front as substrate. The rapid emergence prevented it from growing wide.

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## REFERENCES

ADEY, W.H. (1986) Coralline algae as indicators of sea-level. In: *Sea-level Research: a Manual for the Collection and Evaluation of Data* (Ed. van de Plassche, O.), pp. 229–279. Geobooks, Norwich.

Adey, W.H., Townsend, R.A. & Boykins, W.T. (1982) The crustose coralline algae (Rhodophyta: Corallinaceae) of the Hawaiian Islands. *Smithson. Contrib. mar. Sci.*, **15**, 1–74.

Aissaoui, D.M. (1988) Magnesian calcite cements and their diagenesis: dissolution and dolomitization, Mururoa atoll. *Sedimentology*, **35**, 821–841.

Aissaoui, D.M. & Purser, B.H. (1986) La cimentation dans les récifs: principe de cimentation maximale. *C. R. Acad. Sci. Paris*, **303**, 301–303.

Bard, E., Hamelin, B., Fairbanks, R.G. & Zindler, A. (1990) Calibration of the $^{14}$C timescale over the past 30 000 years using mass spectrometric U–Th ages from Barbados corals. *Nature*, **345**, 405–410.

Bard, E., Arnold, M., Fairbanks, R.G. & Hamelin, B. (1993) $^{230}$Th–$^{234}$U and $^{14}$C ages obtained by mass spectrometry on corals. *Radiocarbon*, **35**, 191–199.

Bard, E., Hamelin, B., Arnold, M. *et al.* (1996) Deglacial sea level record from Tahiti corals and the timing of global meltwater discharge. *Nature*, **382**, 241–244.

Bathurst, R.G.C. (1966) Boring algae, micrite envelopes and lithiflcation of molluscan biosparites. *Geol. J.*, **5**, 15–32.

Beck, J.W., Edwards, R.L., Ito, E. *et al.* (1992) Sea surface temperature from coral skeletal strontium/calcium ratios. *Science*, **257**, 644–647.

Beck, J.W., Récy, J., Taylor, F.W., Edwards, R.L. & Cabioch, G. (1997) Holocene sea surface temperature variations in the SW Pacific Ocean derived from coral Sr/Ca and oxygen isotope thermometry. *Nature*, **385**, 705–707.

Benoit, M. & Dubois, J. (1971) The earthquake swarm in the New Hebrides archipelago, August 1965. *R. Soc. N. Z.*, **9**, 141–148.

Bloom, A.L. & Yonekura, N. (1985) Coastal terraces generated by sea-level change and tectonic uplift. In: *Models in Geomorphology* (Ed. Woldenberg, M.J.), pp. 139–154. Allen and Unwin, Winchester, MA.

Bloom, A.L., Broecker, W.S., Chappell, J.M.A., Matthews, R.K. & Mesolella, K.J. (1974) Quaternary sea-level fluctuations on a tectonic coast: new Th230/U234 dates from the Huon Peninsula, New Guinea. *Quat. Res.*, **4**, 185–205.

Bosence, D. (1984) Construction and preservation of two modern coralline algal reefs, Saint-Croix, Caribbean. *Palaeontology*, **27**, 549–574.

Bosence, D. (1985) Preservation of coralline–algal reef frameworks. In: *Proceedings of the 5th International Coral Reef Congr.*, *Tahiti*, **6**, pp. 623–628.

Cabioch, G. (1988) *Récifs frangeants de Nouvelle-Calédonie (Pacifique sud-ouest). Structure interne et influences de l'eustatisme et de la néotectonique.* Thèse Doct., Université Aix–Marseille I.

Cabioch, G., Montaggioni, L.F. & Faure, G. (1995) Holocene initiation and development of New Caledonian fringing reefs, South-West Pacific. *Coral Reefs*, **14**, 131–140.

Camoin, G.F. & Montaggioni, L.F. (1994) High energy coralgal–stromatolite frameworks from Holocene reefs (Tahiti, French Polynesia). *Sedimentology*, **41**, 655–676.

Chappell, J. (1974) Geology of coral terraces, Huon Peninsula, New Guinea: a study of Quaternary tectonic movements and sea-level changes. *Geol. Soc. Am. Bull.*, **85**, 553–570.

Chappell, J. & Polach, H. (1991) Post-glacial sea-level rise from a coral record at Huon Peninsula, Papua New Guinea. *Nature*, **349**, 147–149.

Chappell, J., Ota, Y. & Berryman, K.R. (1994) Holocene and Late Pleistocene coseismic uplift of Huon Peninsula, Papua New Guinea. In: *Study on Coral Reef Terraces of the Huon Peninsula, Papua New Guinea* (Ed. Ota, Y.), pp. 35–62. Department of Geography, Senshu University, Kawasaki, Japan.

Collins, L.B., Zhu, Z.R., Wyrwoll, K.-H. *et al.* (1993) Holocene growth history of a reef complex on a cool-water carbonate margin: Easter Group of the Houtman Abrolhos, Eastern Indian Ocean. *Mar. Geol.*, **115**, 29–46.

Davies, P.J. & Hopley, D. (1983) Growth facies and growth rates of Holocene reefs in the Great Barrier Reef. *Bur. Min. Res. J. Aust. Geol. Geophys.*, **8**, 237–251.

Davies, P.J. & Marshall, J.F. (1979) Aspects of Holoçene reef growth—substrate age and accretion rate. *Search*, **10**, 276–279.

Davies, P.J. & Montaggioni, L.F. (1985) Reef growth and sea-level change: the environmental signature. In: *Proceedings of the 5th International Coral Reef Congr.*, *Tahiti*, **3**, pp. 477–511.

Davies, P.J., Marshall, J.F. & Hopley, D. (1985) Relationships between reef growth and sea level in the Great Barrier Reef. In: *Proceedings of the 5th International Coral Reef Congr.*, *Tahiti*, **3**, pp. 95–103.

Done, T.J. (1982) Patterns in the distribution of coral communities across the central Great Barrier Reef. *Coral Reefs*, **1**, 95–107.

Done, T.J. & Navin, K.F. (1990) Shallow-water benthic communities on coral reefs. In: *Vanuatu Marine Resources: Report of a Biological Survey* (Ed. Done, T.J. & Navin, K.F.), pp. 10–36. Australian Institute of Marine Science, Townsville, Qld.

Easton, W.H. & Olson, E.A. (1976) Radiocarbon profile of Hanauma Reef, Oahu, Hawaii. *Geol. Soc. Am. Bull.*, **87**, 711–719.

Edwards, R.L., Chen, J.H. & Wasserburg, G.J. (1987) $^{238}$U–$^{234}$U–$^{230}$Th–$^{232}$Th systematics and the precise measurement of time over the past 500 000 years. *Earth planet. Sci. Lett.*, **81**, 175–192.

Edwards, R.L., Beck, W.J., Burr, G.S. *et al.* (1993) A large drop in atmospheric $^{14}$C/$^{12}$C reduced melting in the Younger Dryas, documented with $^{230}$Th ages of corals. *Science*, **260**, 962–968.

Fairbanks, R.G. (1989) A 17 000-year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. *Nature*, **342**, 637–642.

Faure, G. (1982) *Recherche sur les peuplements des scléractiniaires des récifs coralliens de l'archipel des Mascareignes (Océan Indien occidental).* Thèse Doct. ès Sci., Université Aix–Marseille II.

Gilpin, L. (1982) *Tectonic geomorphology of Santo Island, Vanuatu.* MS thesis, Cornell University, Ithaca, NY.

Gray, S.C., Burr, G.S., Donahue, D.J. *et al.* (1993) Comparison of TIMS $^{230}$Th and AMS $^{14}$C ages of Pacific corals for the last deglaciation: evidence for a rapid oscillation of delta $^{14}$C. *EOS Trans., Am. geophys. Union, 1993 Fall meeting*, **74**, pp. 326.

Guilcher, A. (1974) Coral reefs of the New Hebrides, Melanesia, with particular reference to open-sea, not fringing, reefs. In: *Proceedings of the 2nd International Coral Reef Symposium, Brisbane, Qld*, **2**, pp. 523–535.

Gvirtzman, G. & Friedman, G.M. (1977) Sequence of progressive diagenesis in coral reefs. *Am. Assoc. petrol. Geol., Stud. Geol.*, **4**, 357–380.

Henrich, R. & Wefer, G. (1986) Dissolution of biogenic carbonates: effects of skeletal structures. *Mar. Geol.*, **71**, 341–362.

Hopley, D. (1982) *The Geomorphology of the Great Barrier Reef: Quaternary Development of Coral Reefs.* John Wiley, New York.

James, N.P. (1974) Diagenesis of scleractinian corals in the subaerial vadose environment. *J. Paleontol.*, **48**, 785–799.

Johnson, D.P. & Risk, M.J. (1987) Fringing reef sequences and geochemistry, Britomart Reef, central Great Barrier Reef, Australia. *Sedimentotogy*, **31**, 515–529.

Jouannic, C., Taylor, F.W., Bloom, A.L. & Bernat, M. (1980) Late Quaternary uplift history from emerged reef terraces on Santo and Malekula, Central New Hebrides island arc. *UN ESCAP, CCOP/SOPAC, Tech. Bull.*, **3**, 91–108.

Kan, H., Nori, N., Nakashima, Y. & Ichikawa, K. (1995) The evolution of narrow reef flats at high-latitude in the Ryukyu Islands. *Coral Reefs*, **14**, 123–130.

Lambeck, K. & Nakiboglu, S.M. (1986) Recent sea level changes in the south-west Pacific. *R. Soc. N. Z. Bull.*, **24**, 485–491.

Larcombe, P., Carter, R.M., Dye, J., Gagan, M.K. & Johnson, D.P. (1995) New evidence for episodic postglacial sea-level rise, central Great Barrier Reef, Australia. *Mar. Geol.*, **127**, 1–44.

Littler, M.M. & Doty, M.S. (1975) Ecological components structuring the seaward edges of tropical Pacific reefs: the distribution communities and productivity of *Porolithon. J. Ecol.*, **63**, 117–129.

Macintyre, I.G. (1977) Distribution of submarine cements in a modern Caribbean fringing reef, Galeta Point, Panama. *J. sediment. Petrol.*, **47**, 503–516.

Macintyre, I.G. & Marshall, J.F. (1988) Submarine lithification in coral reefs: some facts and misconceptions. In: *Proceedings of 6th International Coral Reef Symposium, Townsville, Qld*, **1**, pp. 263–272.

Marshall, J.F. (1986) Regional distribution of submarine cements within an epicontinental reef system: central Great Barrier Reef, Australia. In: *Reef Diagenesis* (Eds Schroeder, J.H. & Purser, B.H.), pp. 8–26. Springer-Verlag, Berlin.

Marshall, J.F. & Davies, P.J. (1982) Internal structure and Holocene evolution of One Tree Reef, Southern Great Barrier Reef. *Coral Reefs*, **1**, 21–28.

McCulloch, M., Mortimer, G., Esat, T., Xianhua, L., Pillans, B. & Chappell, J. (1996) High resolution windows into early Holocene climate: Sr/Ca coral records from the Huon Peninsula. *Earth planet. Sci. Lett.*, **138**, 169–178.

Montaggioni, L.F. (1988) Holocene reef growth history in mid-plate high volcanic islands. In: *Proceedings of the 6th International Coral Reef Symposium, Townsville, Qld*, **3**, pp. 455–460.

Montaggioni, L.F. & Camoin, G.F. (1993) Stromatolites associated with coralgal communities in Holocene high-energy reefs. *Geology*, **21**, 149–152.

Morton, J.E. (1973) The intertidal ecology of the British Solomon Islands. 1. Zonation patterns of the weather coasts. *Phil. Trans. R. Soc. London*, **265**, 491–542.

Morton, J.E. (1974) The coral reefs of the British Solomon Islands: a comparative study of their composition and ecology. In: *Proceedings of the 2nd International Coral Reef Symposium, Brisbane, Qld*, **2**, pp. 31–53.

Morton, J.E. & Challis, D.A. (1969) The biomorphology of Solomon Islands shores with a discussion of zoning patterns and ecological terminology. *Phil. Trans. R. Soc. London, Ser.B*, **255**, 459–516.

Neumann, A.C. & Macintyre, I. (1985) Reef response to sea level rise: keep-up, catch-up or give-up. In: *Proceedings of the 5th International Coral Reef Congr., Tahiti*, **3**, pp. 105–110.

Nunn, P.D. (1993) Role of *Porolithon* algal-ridge growth in the development of the windward coast of Tongatapu island, Tonga, South Pacific. *Earth Surf. Process. Landf.*, **18**, 427–439.

Ota, Y. & Chappell, J. (1996) Late Quaternary coseismic uplift events on the Huon Peninsula, Papua New Guinea, deduced from coral terrace data. *J. geophys. Res.*, **101**, 6071–6082.

Ota, Y., Chappell, J., Kelley, R. *et al.* (1993) Holocene coral reef terraces and coseismic uplift of Huon Peninsula, Papua New Guinea. *Quat. Res.*, **40**, 177–188.

Partain, B.R. & Hopley, D. (1989) *Morphology and development of the Cape Tribulation fringing reefs, Great Barrier Reef Australia.* Great Barrier Reef Marine Park Authority Tech. Mem., **21**, 1–45.

Pirazzoli, P.A. (1986) Marine notches. In: *Sea-level Research: a Manual for the Collection and Evaluation of Data* (Ed. van de Plassche, O), pp. 361–400. Geobooks, Norwich.

Récy, J., Beck, J.W., Taylor, F.W., Edwards, R.L. & Cabioch, G. (1993) Variations holocènes de la température de surface de la mer dans le Sud-Ouest Pacifique. *Réun. spéc. Soc. géol. Fr.* Soc. géol. Fr., Paris, 90 pp.

Richard, G. (1982) *Mollusques lagunaires et récifaux de Polynésie fançaise. Inventaire faunistique, bionomie, bilan quantitatif croissance, production.* Thèse Doct. ès Sci., Université P. & M. Curie, Paris.

Stuiver, M. & Reimer, P.J. (1993) Extended $^{14}$C data base and revised calib 3.0 $^{14}$C age calibration program. *Radiocarbon*, **35**, 215–230.

Taylor, F.W. (1992) Quaternary vertical movements of the central New Hebrides island arc. In: *Proceedings of the Ocean Drilling Program, Initial Reports*, 134 (Eds Collot, J.-Y, Greene, H.G. & Stokking, L.B.), pp. 33–42. Ocean Drilling Program, College Station, TX.

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

Taylor, F.W., Jouannic, C. & Bloom, A.L. (1985) Quaternary uplift of the Torres Islands, northern New Hebrides frontal arc: comparison with Santo and

Malekula islands, central New Hebrides frontal arc. *J. Geol.*, **93**, 419–438.

TAYLOR, F.W., FROHLICH, C., LECOLLE, J. & STRECKER, M. (1987) Analysis of partially emerged corals and reef terraces in the central Vanuatu arc: comparison of contemporary coseismic and nonseismic with Quaternary vertical movements. *J. geophys. Res.*, **92**, 4905–4933.

TAYLOR, F.W., EDWARDS, R.L. & WASSERBURG, G.J. (1990) Seismic recurrence intervals and timing of aseismic subduction inferred from emerged corals and reefs of the central Vanuatu (New Hebrides) frontal arc. *J. geophys. Res.*, **95**, 393–408.

TAYLOR, F.W., MANN, P., LAGOE, M. *et al.* (1993) Mechanisms for rapid reversals of vertical motion in the New Hebrides and Solomon arcs related to colliding features. *EOS Trans., Am. geophys. Union, 1993 Fall Meeting*, **74**, 545 pp.

VERON, J.E.N. (1990) Checklist of the hermatypic corals of Vanuatu. *Pacific Sci.*, **44**, 51–70.

WALBRAN, P.D. (1994) The nature of the pre-Holocene surface, John Brewer Reef, with implications for the interpretation of Holocene reef development. *Mar. Geol.*, **122**, 63–79.