



# History of Indo-Pacific coral reef systems since the last glaciation: Development patterns and controlling factors

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Received 25 May 2004; accepted 10 January 2005

## Abstract

A significant body of new information about the development of coral reefs during the last 23 ka has been generated in the last three decades. In the Indo-Pacific province, structures from a variety of geodynamic settings have been investigated using subsurface drilling and submersible diving. This paper is based principally on the re-examination of the core dataset from the literature, with reconversion of many previously published radiocarbon ages into calendar dates.

Seven framework and three detrital facies were identified on the basis of the nature and growth shapes of dominant framework builders, and on that of the texture of sediments, respectively. Framework facies in high-hydrodynamic energy settings were dominated by an association of coralline algae and robust-branching corals (*Acropora robusta* group, *A. gr. humilis*, *A. palifera*, *Pocillopora damicornis*) with locally encrusting coral forms (faviids). In moderate energy environments, these were replaced by domal (*Porites*), tabular-branching (*Acropora gr. hyacinthus*) and arborescent (*Acropora gr. muricata*), whereas sheltered areas included an association of arborescent, foliaceous (*Montipora*, *Pavona*) and encrusting coral species. Detrital facies comprise coral rubble, carbonate sand and mud. On compositional and textural bases, four main sand subfacies were recognized: coralgal rudstone to packstone; coral–molluscan grainstone/packstone; molluscan–foraminiferal grainstone/packstone; and green algal (*Halimeda*) grainstone/packstone. Despite some overlaps in the sand facies association, each subfacies can provide additional support to reconstruction of paleoreef environments.

Three types of framework facies association were identified within entire reef-margin sequences: framework of homogeneous composition reflecting stability of environmental conditions through time; superimposition of two distinct frameworks, usually as deeper water corals overlain by shallower, higher energy ones, and recurrent alternations of shallower and deeper coral assemblages. The two last associations resulted probably from lateral displacements of coral communities in response to rapid changes in accommodation space. Such facies transitions also are described from backreef sediment piles: gravel graded into sand and mud successively as a result of upward shallowing. The degree of reef development seems to be linked to coral community structure. Communities consisting principally of branching and domal coral forms favoured substantial accretion and the formation of well-developed reefs, whereas assemblages comprising foliaceous and encrusting colonies produced only incipient reefs. Within reef systems, the proportions of detritus over framework tend to increase as

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hydrodynamic energy declines. The Indo-Pacific reef systems are classified into four anatomy types on the basis of dominant depositional patterns: balanced aggrading/onlapping, unbalanced aggrading/downlapping, prograding and backstepping types. Vertical accretion rates of frameworks are highly variable and are not directly dictated by coral growth habits. However, the highest rates recorded (up to 20 mm year<sup>-1</sup>) relate to tabular- and arborescent-acroporid rich sections. Abrupt variations in the aggradation rates of framework are recorded in sequences at the transitional zone between two distinct coral assemblages. In detritus-dominated sequences, accumulation rates range from 0.2 to about 40 mm year<sup>-1</sup>, with higher values suggesting intense hurricane-controlled deposition. In addition, accretion rates also seem to depend on water-energy conditions. In high-energy environments, aggradation rates did not exceed 12 mm year<sup>-1</sup>, but reached 25 mm year<sup>-1</sup> in more protected areas. By contrast, lateral accretion operated at an average rate of 90 mm year<sup>-1</sup> in agitated waters, while it did not exceed the mean rate of 55 mm year<sup>-1</sup> in calm waters. Changes in accretion rates appear to be linked to reef growth modes. In the reef zones driven by a “keep-up” mode, mean vertical accretion rates range at around 6 mm year<sup>-1</sup>. The reef zones developed through a “catch-up” mode at rates of 3–4 mm year<sup>-1</sup>. There was little variation in accretion rates according to latitude.

At the Last Glacial Maximum, from 23 to about 19 ka BP, reefs (Reef Generation RGO) only developed along what were to become the foreslopes of present reefs, forming accumulations a few metres thick at vertical rates of up to 1 mm year<sup>-1</sup>. The rapid postglacial rise in sea level, from about 19 to 6.5 ka BP, was accompanied by the settlement of three successive reef generations (the so called RGI, RGII and RGIII), within the periods 17.5–14.7, 13.8–11.5 and 10 ka BP to the Present.

During the Postglacial transgression, regional to local differences in gross morphology and internal architecture of the reefs have been determined by differing sea-level histories in combination with neotectonics and topographic factors. Locally, reef colonization seems to have been facilitated or prevented chiefly by small-scale topographic features. Development during subsequent deglaciation was probably largely independent of variations in sea surface temperatures. Water turbidity also seems to have been only a minor determinant of reef settlement and growth, but may locally have controlled the composition of coral communities, resulting in the growth of turbidity-tolerant domal and foliaceous forms.

Changes in atmospheric CO<sub>2</sub> levels remained within the tolerance thresholds for reef calcification. The three main reef growth episodes coincide roughly with rapid increases in atmospheric *p*CO<sub>2</sub>. Dust input and variations in sea surface salinities seem to have had a very limited control on reef growth. The LGM was characterized by salinities comparable with those of the present, but by higher dust fluxes. By contrast, nutrient levels, hydrodynamic energy, and to a lesser extent coral recruitment in relation to substrate availability and ocean circulation, have played major roles in determining reef accretion patterns at both local and regional scales. Two periods of increased upwelling in the western Indian Ocean, at 15.3 and 11.5–10.8 ka BP, coincided with the demise of RGI and RGII. During deglaciation, high-frequency storm events probably led to a scarcity of typical growth framework reefs and favoured the formation of structures composed of reworked and recemented coral framework. Storm control may have been particularly important in the mid-Holocene when water depths over incipient reefs were greater than 5 m. From the LGM to the early Holocene, coral settlement has probably declined due to a lack of suitable nurseries, until the modern patterns of ocean circulation were established and thus favoured larval dispersal from refuges. It is highly desirable to improve analysis of the core database and to increase the number of core-transects, including forereef sites, to enhance our knowledge of Recent reef development.

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*Keywords:* Corals; Reefs; Growth; Paleoenvironments; Indian Ocean; Pacific Ocean; late Pleistocene; Holocene

## 1. Introduction

A long-standing objective of carbonate geologists is to understand the structure and composition of organic reefs and to learn what were the main physical–chemical forcing functions that controlled reef development through time and how they interacted. Recent reefs therefore have been used for a long time as counterparts for interpreting ancient reefs, even where the origin, growth patterns and controlling environ-

mental factors of Holocene reefs were still poorly documented. At present, the history of Holocene reefs from the tropical Atlantic province is far better understood than that of the Indo-Pacific areas, probably because of two major reasons: (1) the tropical Indo-Pacific seas occupy a very extensive area (130 millions km<sup>2</sup>), i.e. 25 times greater than the Caribbean and adjacent tropical waters; (2) the zonation and community structure of Atlantic reefs are apparently simpler than those observed in the Indo-Pacific.

The scientific investigation of Indo-Pacific coral reefs began with the visits of von Chamisso and von Kutzebue (1815–1819), Quoy and Gaimard (1817–1818), and Beechy (1825–1828), summarized by Lyell (1832). Darwin (1842) was the first to propose a reasonable (subsidence-controlled) model of coral reef formation based on observations of the morphology of Polynesian islands. This was the starting point for over a century of scientific controversy. In particular, Daly (1910) contested Darwin's conclusions and suggested that the foundations of present-day reefs resulted from marine erosion during Pleistocene low sea-level stands, that is, during glacial periods (reviews by Stoddart, 1969; Steers and Stoddart, 1977; Hopley, 1982, pp. 1–25).

From early in the 20th century, it was clear that the Indo-Pacific area was destined to be a focus of reef research. The voyage of Agassiz (1903) in the tropical Pacific Ocean and two Indian Ocean expeditions of Gardiner (1898) dramatically extended knowledge of reef geology. As early as 1896, the Royal Society in Britain together with Australian geologists began deep drilling on Funafuti Atoll in search of reef foundations (David and Sweet, 1904; Royal Society, 1904). Deep coring through reefs subsequently took place on Kita-Daito-Jima Atoll (Ota, 1938, in Ladd et al., 1970), on the Great Barrier Reef of Australia (Richards and Hill, 1942), in the Marshall islands (Emery et al., 1954; Ladd and Schlanger, 1960), on Midway Atoll (Ladd et al., 1970), in New Caledonia (Avias and Coudray, 1967) and on Mururoa Atoll (Deneufbourg, 1969; Repellin, 1975). These works broadly supported Darwin's subsidence theory but also restated Daly's theory of glacial control on reef morphology in terms of karstic processes (Steers and Stoddart, 1977; Szabo et al., 1985; Buigues et al., 1992; Purdy and Winterer, 2001). However, as emphasized by Steers and Stoddart (1977), while deep coring increased knowledge of the Tertiary history of a number of reefs, only rarely did it provide a thorough understanding of the geology and structure of Postglacial reef sections. Most studies of Recent coral reefs have focused on physiography, the distribution of biota and surface sedimentation. Principally descriptive, they have delivered little information on recent geological history (Braithwaite, 1982a,b).

The development of light-weight drilling rigs (Thom, 1978) opened a new era of shallow coring

world-wide from the 1970s onwards. New data relating to the interior structure and growth history of Holocene reefs was obtained from a variety of geodynamic settings. Accordingly, new reef models emerged, focusing in particular on the relationship between sea-level changes and reef growth (Macintyre and Glynn, 1976; Lighty et al., 1982; Hopley, 1982; Davies et al., 1985; Davies and Montaggioni, 1985; Neumann and Macintyre, 1985; Macintyre, 1988; Montaggioni, 1988).

In the Indo-Pacific province, the starting phase of subsurface reef coring was undertaken in oceanic islands: Tahiti (Deneufbourg, 1971), Marshall islands (Henny et al., 1974), Réunion (Montaggioni, 1976, 1977), Hawaiian islands (Easton and Olson, 1976). Intensive programmes of shallow drilling through continental shelf reef systems commenced from 1973 with the studies on the Great Barrier Reef of Australia (Davies et al., 1977; Hopley, 1977). The earliest shallow drilling operation on Holocene reefs developed on island arcs was conducted by Konishi et al. (1978, 1983) in the Ryukyus.

The two following decades were highly productive for reconstructing the three-dimensional development histories of the Indo-Pacific reefs. Outstanding results have been obtained on the Holocene evolution of continental shelf reefs owing to the high drilling capacity of Australian teams; between 1973 and 1990, about 140 holes were drilled into 44 reefs throughout the Great Barrier Reef (Hopley and Davies, submitted for publication). Although drilling operations through Recent mid-oceanic and forearc reefs were scattered, the knowledge of their development has increased significantly. All in all, locally enriched by results from submersible diving along reef foreslopes, an important databank has been provided, regarding the timing of Holocene reef establishment, the internal architecture and facies relationship and accretion rates. The major findings gained at that time can be summarized as follows: (1) widespread occurrence of Last Glacial reef systems, at present submerged; (2) identification of at least four successive reef generations since the Last Glacial Maximum; (3) growth commencement of most of the modern reefs from 9.5 to 7 ka BP, commonly from karstified pre-Holocene foundations; (4) identification of three basic responses of reef growth to sea-level rise: "keep-up" mode referring to reefs initiated as soon as foundations were

flooded and developed at the same rate as sea-level rise, “catch-up” mode typifying reefs that displayed a lag in initiation and then caught up with sea level, and “give-up” mode defining reefs that failed to reach sea level and were drowned; (5) identification of the composition of corallgal communities and of their significance in terms of paleobathymetry; (6) control of internal facies by variations in water energy; (7) dominance of lateral accretion after sea-level stabilization between 6.5 and 3 ka BP; (8) regional changes in reef flat ages; (9) reconstruction of Pacific sea-level curves for the past 14 ka BP based on corrected reef growth curves; (10) minor influence of sea surface temperature on reef development during the Glacial and Postglacial times.

With the increasing realisation that coral reefs are able to provide reliable records of paleoenvironmental changes (see, for instance, [Montaggioni and Macintyre, 1991](#); [Quinn and Tudhope, 2000](#)), reef geologists tried to integrate neocological studies conducted by biologists (see [Cornell and Karlson, 2000](#)) on the effects of environmental variables on coral and reef growth. This has helped with understanding of the respective role of physical–chemical parameters in the development of Holocene reefs. [Hubbard \(1988\)](#), [Smith and Buddemeier \(1992\)](#), [Hopley \(1994\)](#), [Brown \(1997\)](#), [Kleypas \(1997\)](#) and [Montaggioni \(2000\)](#) reviewed the major controls on recent reef development.

Comprehensive reviews were published previously on the Postglacial history of Indo-Pacific reefs, but their impact was limited to regional scales: Great Barrier Reef of Australia ([Hopley, 1982, 1994, 1997](#); [Davies and Hopley, 1983](#); [Hopley and Davies, submitted for publication](#)), high volcanic islands ([Glynn and Wellington, 1983](#); [Montaggioni, 1988](#)) and island arcs ([Cabiocch, 2003](#); [Yamano et al., submitted for publication](#)). Apart from recent synthesizing reports based on the analysis of a few reefs around the world ([Hubbard, 1997](#); [Hubbard et al., 1998](#); [Montaggioni, 2000](#); [Hallock, 2001](#)), the only review referring mainly to case studies in the Indo-Pacific province is that devoted to Holocene fringing reef growth by [Kennedy and Woodroffe \(2002\)](#).

The aims of this paper are to describe the development of Indo-Pacific reefs ([Fig. 1](#)) during and after the Last Glacial Maximum (LGM) which spans, over the past 23 ka (ka=1000 years), the major

climatic transition from full glacial, low sea stand to interglacial, high sea stand. It is based on the re-analysis of previously published core data which have been collated for the first time. The paper is an attempt to determine patterns of reef development and to identify the main environmental factors governing that development. Finally, the paper expresses something of the state of knowledge on the subject at the start of the new millennium.

## 2. Materials and methods

### 2.1. *The core database: value and limitations*

Bio-lithological descriptions of the 684 subsurface cores that have penetrated through modern Indo-Pacific reef are reviewed. The data are compiled in [Table 1](#) by geographical region, clockwise from the far Western Indian to Far Eastern Pacific oceans. This review reports the location of subsurface drilling sites, the number of cores extracted from each site, the number of radiometric dates obtained from cored material, the minimum to maximum thickness of the Postglacial reef sections penetrated and the age of the reef base (i.e. age of reef initiation). It intends to provide researchers with a comprehensive database with which to investigate reef growth in the Indo-Pacific.

Diamond drilling and vibrocore techniques have been used to sample the internal parts of the reefs. The core diameters vary between 25 and 85 mm. Except in about ten sites where single or scattered holes only have been made, drilling operations have generally been conducted along core-transects, crossing the major top-reef zones, from the reef crest to the backreef. However, all in all, the number of cores recovered from the reef flats was higher than that extracted from the lagoons.

Little is known about the geological development of reef-front and fore-reef zones in the Indo-Pacific, due to the difficulty in coring offshore. The available data come from drilling through an ocean-facing forereef of Western Australia ([Collins et al., 2003](#)), and front slopes of various inner shelf fringing reefs (Iki island, Japan: [Yamano et al., 2001b](#); Hayman island, Eastern Australia: [Hopley et al., 1983](#); Hawaii: [Grossman and Fletcher, 2004](#); [Engels et al., 2004](#); Gulf of Chiriqui, Panama: [Glynn and Macintyre,](#)

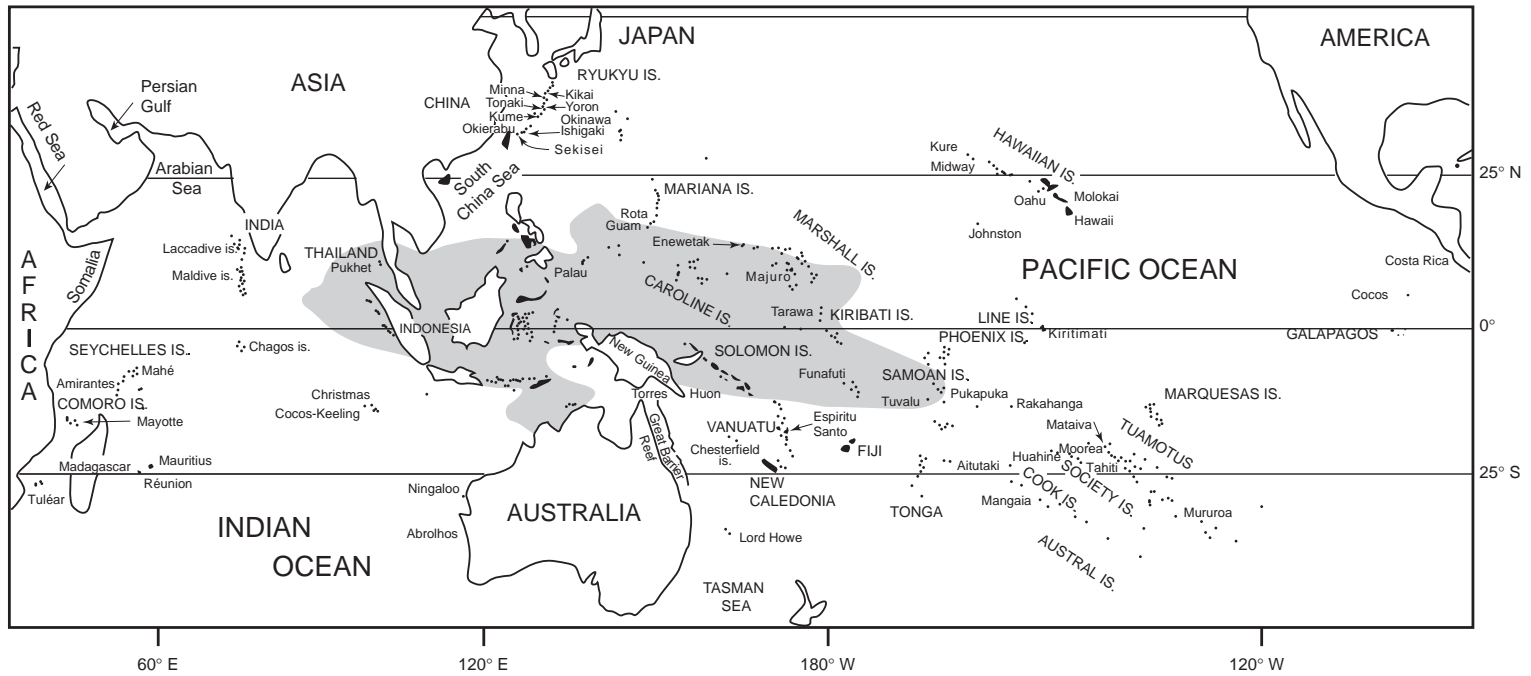


Fig. 1. Index map of the Indo-Pacific province showing the location of the main reef sites mentioned in the present study. The dark area refers to the Indo-Pacific Warm Pool.

Table 1  
Summary of results from shallow drilling operations through modern coral reefs of the Indo-Pacific province

Site location	Reef type	Number of cores drilled	Number of radiometric dates <sup>a</sup>	Thickness range of Postglacial/Holocene reef sections (in m)	Age range of reef initiation (in 1000 calendar years BP)	Source
Western Indian Ocean						
Réunion	fr	1	19	>18	7.72–8.00	Montaggioni (1976, 1977, 1988), Camoin et al. (1997)
Mauritius	fr	3	39	12.8–19.3	6.7–9.60	Montaggioni (1981,1988), Camoin et al. (1997)
Mayotte	br/fr	14	58	2.9–21.5	2.25–9.72	Camoin et al. (1997), Zinke et al. (2003a,b)
Mahé, Seychelles	fr	8	16	>26	5.43–10.59	Braithwaite et al. (2000)
Toliara, Madagascar	br	2	19	12.9–13.4	6.63–8.34	Camoin et al. (2004)
Red Sea						
Egypt	pr	4	NA	>15	NA	Fletcher et al. (2000)
Eastern Indian Ocean						
Maldives, Mahé	alt	1,pits	7	>17.2	>6.87	Woodroffe (1992)
Cocos–Keeling	alt	1,pits	15	7→15	6.72	Woodroffe (1994), Woodroffe et al. (1999)
Pukhet, Thailand	fr	21	12	4→4	0.55–6.62	Tudhope and Scoffin (1994)
Abrolhos, SW Australia	pr	6	18	4→25	7.09–9.80	Collins et al. (1993), Eisenhauer et al. (1993)
Ningaloo, SW Australia	fr	12	8	18	4.37–7.57	Collins et al. (2003)
Western Pacific Ocean						
Ryukyu islands						
Iki	fr	3	5	>3	>1.40	Yamano et al. (2001b)
Kikai-Jima	fr	24 exc	69	>3–25	8.95–10.1	Webster et al. (1998), Yamano et al. (submitted for publication)
Okierabu-Jima	fr	4 exc	30	>2–11	7.5–7.9	Kan et al. (1995), Ichikawa et al. (1994), Yamano et al. (submitted for publication)
Yoron-tou	fr	4	10	>4	>6.1	Yonekura et al. (1994), Yamano et al. (submitted for publication)
Okinawa-Hontou	fr	25	21	10–20	7.57–8.53	Yamano et al. (submitted for publication)
Minna-Jima	fr	exc	20	>3	>6.0	Yamano et al. (submitted for publication)
Tonaki-Jima	fr	exc	25	>6	>6.30	Kan et al. (1997a,b,c)
Kume-Jima	fr	9	37	6–11	7.72–8.23	Takahashi et al. (1988), Kan et al. (1991), Yamano et al. (submitted for publication)
Ishigaki-Jima	fr	17 exc	56	>5–22	>6.40–9.42	Yamano et al. (2001a, submitted for publication)
Sekisei area	fr	3	NA	17–21.5	8.45	Yamano et al. (submitted for publication), Kawana and Kan (2000)
Eastern Australia Great Barrier Reef <sup>b</sup>						
Northern Region (10–17°S)	fr/pr/br	36	84	3.9–25.2	5.60–10.30	Hopley and Davies (submitted for publication)



Table 1 (continued)

Site location	Reef type	Number of cores drilled	Number of radiometric dates <sup>a</sup>	Thickness range of Postglacial/Holocene reef sections (in m)	Age range of reef initiation (in 1000 calendar years BP)	Source
Eastern Australia Great Barrier Reef <sup>b</sup>						
Central Region (17–20°S)	fr/pr/br	52	304	1–30	5.48–9.95	Hopley and Davies (submitted for publication)
Southern Region (20–24°S)	fr/pr/br	72	231	8.1–14	6.38–8.52	Hopley and Davies (submitted for publication), Smith et al. (1998)
Torres Strait Islands	pr	17	27	4.5–8.5	7.57–7.92	Woodroffe et al. (2000)
Lord Howe Island	fr	19	55	7–23	6.79–>7.22	Kennedy and Woodroffe (2000)
Middleton and Elisabeth reefs	pr	25	27	8–>40	6.7–>6.7	Woodroffe et al. (2004)
New Caledonia	fr	37	33	3–13.5	2.00–8.05	Cabioch (1988, 2003), Cabioch et al. (1995, 1999c)
	br	13	NA	10–15	7.50–9.00	Coudray (1976), Cabioch (2003)
	atl	3	7	7–7.9	NA	Degaugue-Michalski (1993)
Chesterfield islands	atl	13	16	>14.5	>7.29	Degaugue-Michalski (1993)
Huon Peninsula, Papua New Guinea	fr	3	47	50–<58.7	23.6	Ota and Chappell (1999), Edwards et al. (1993), Cutler et al. (2003)
Vanuatu, Espiritu Santo	fr	18	43	30–37	20.5–24.3	Cabioch et al. (1998, 2003a), Cabioch (2003)
Mariana islands						
Rota	fr	5	14	>7	>6.11	Kayanne et al. (1993)
Guam	fr	4	12	5.5–>5.8	>3.88	Kayanne et al. (1993)
Palau islands	br	6	21	14.5–24	7.97–8.31	Kayanne et al. (2002)
Marshall islands						
Bikini	atl	3	1	8–10	>6.17	Emery et al. (1954), Tracey and Ladd (1974)
Enewetak	atl	11	3	7–12	7.90–8.20	Ladd and Schlanger (1960), Tracey and Ladd (1974), Szabo et al. (1985)
Kiribati islands						
Tarawa	atl	10	22	10.6–16.5	7.60–9.01	Marshall and Jacobsen (1985)
Tuvalu islands						
Funafuti	atl	1	9	26.4–27.4	8.1	Ohde et al. (2002)
Line islands						
Kiritimati	atl	24	NA	2–5.3	<4.5	Woodroffe and Mc Lean (1998)
Central Pacific Ocean						
Hawaiian islands						
Oahu	fr	47	98	9.5–18	7.93–7.49	Easton and Olson (1976), Grigg (1998), Grossman and Fletcher (2004)
Midway	br	2	NA	3–5	NA	Ladd et al. (1970)
Molokai	fr	24	31	>5.3	>8.14	Engels et al. (2004)
Cook Islands						
Pukapuka	atl	3	5	14.5–22.0	9.2	Gray et al. (1992)
Rakahanga	atl	2	3	15–17	>8.4	Gray et al. (1992)
Mangaia	fr	5	8	>5	>5.29	Yonekura et al. (1988)

(continued on next page)

Table 1 (continued)

Site location	Reef type	Number of cores drilled	Number of radiometric dates <sup>a</sup>	Thickness range of Postglacial/Holocene reef sections (in m)	Age range of reef initiation (in 1000 calendar years BP)	Source
Central Pacific Ocean						
Society islands						
Tahiti	fr/br	14	80	52–87	13.8	Deneufbourg (1971), Montaggioni (1988), Bard et al. (1996), Montaggioni et al. (1997a,b), Cabioch et al. (1999a) Montaggioni (1988)
Moorea	fr/br	3	15	>25	>5.73	
Tuamotu islands						
Mataiva	atl	4	9	4–8.8	6.71	Pirazzoli and Montaggioni (1986)
Mururoa	atl	10	14	3–24	8.3–9.0	Lalou et al. (1966), Repellin (1975), Perrin (1989), Buigues et al. (1992), Camoin et al. (2001)
Eastern Pacific Ocean						
Galapagos,						
Isabella island	fr	4 exc	2	0.5–>2	<1.0	Macintyre et al. (1992)
Costa Rica						
Punta isolates	fr	13	26	2–9	0.94–5.70	Macintyre et al. (1992), Cortés et al. (1994)
Cocos island	fr	6	2	3–3.10.	25–0.43	Macintyre et al. (1992)
Cano island	fr	11 exc	5	0.5–<1	0.24–1.52	Macintyre et al. (1992)
Panama,	fr	12 exc	26	3–>13.4	4.74–5.94	Glynn and Macintyre (1987), Macintyre et al. (1992)
Gulf of Chiriqui						

exc=occurrence of cut excavations and trench walls through reef-front, reef-flat or backreef zones.

Reef type: fr=fringing reef; br=barrier reef; pr=platform reef; atl=atoll.

<sup>a</sup> The compiled dates from reef coring include conventional <sup>14</sup>C and AMS <sup>14</sup>C dates, and <sup>230</sup>Th/<sup>234</sup>U dates determined by alpha spectrometry or by thermal ionization mass spectrometry (TIMS). All of the ages are expressed in calibrated (calendar) years (×1000) BP (cal.ka BP). Dates obtained from samples collected on surficial reef features are not included.

<sup>b</sup> The regional limits within the Great Barrier Reef area are those defined by Maxwell (1968).

1987; Punta Isolates, Costa Rica: Cortés et al., 1994), from underwater observations of excavations or trench walls cut into outer reef slopes (Ryukyu islands, Japan: Yamano et al., 2003, submitted for publication), and from drilling through fore-reef zones or direct observations of seacliffs in tectonically uplifted areas (Huon Peninsula, Papua New Guinea: Ota and Chappell, 1999; Vanuatu: Cabioch et al., 1998).

The drilled sequences range between less than 1 m to up to 87 m in thickness, averaging 10–15 m. About 75% of the drillholes have penetrated the unconformity at the base of the deglacial sequence into the antecedent foundation. Recovery rates have depended upon the type of material penetrated and abundance of

cavities. From framework, recovery range between 60% and 100%, and, from detrital sediments, recovery is 1% to 70%.

Unlike the Caribbean where the anatomy of relic reefs has been well documented (Macintyre, 1972; Macintyre et al., 1981; Fairbanks, 1989; Blanchon and Shaw, 1995; Blanchon et al., 2002), there is little information about the internal structure of Indo-Pacific reefs submerged during the deglacial sea-level rise, even though some of these features have been identified in both oceans in the early 20th century (Davis, 1928: Papua New Guinea, Fidji; Tayama, 1952: Caroline islands, Palau; Madagascar: Guilcher, 1954). Afterwards, research into the location and origin of relic reefs occurring on shelf edges and



foreslopes has been carried out throughout the Indo-Pacific: Madagascar (Pichon, 1978); Seychelles, Chagos (Stoddart, 1971); Maldives (Purdy and Bertram, 1993); Western Indian shelf (Vora et al., 1996); Eastern Australia (Carter and Johnson, 1986; Harris and Davies, 1989); Johnson atoll, Central Pacific (Keating, 1985); Hawaii (Moore et al., 1990; Fletcher and Sherman, 1995; Webster et al., 2003); Marquesas, French Polynesia (Cabioch et al., 2003b). However, there have been few comprehensive studies on these submerged features (see Table 2). The database include results from echo-sounding, seismic profiling, side-scan sonar imaging and bottom photography, combined with dredging, direct submersible observations or Remotely Operated Vehicle (ROV) operations. These have revealed a variety of reef types (mounds, fringing reefs, linear barrier reefs, isolated pinnacles) at depths of 20 to 160 m. Ages derived from surface coral samples indicate that the submerged reefs were active between 20 and 8.5 ka BP. To my knowledge, no drilling has been conducted until now through drowned reefs along the foreslopes of Indo-Pacific shelves. This is due to the fact that collection of coral sections at present submerged have proven difficult. In addition, the value of such a material for reconstructing Glacial to Holocene reef growth history is questionable. As emphasized by Harris and Davies (1989) and Fletcher and Sherman (1995), in most areas, submerged reefs are not typically simple structures developed during the last 23 ka; they may predate the LGM and be features of Pleistocene age, capped by only a thin coralgal veneer deposited during a brief episode of recolonization in the deglacial times.

Figs. 2–4 present a set of 53 core descriptions out of the 684 available, that are grouped according to the reef zones where the corresponding cores came from, i.e. exposed, semi-exposed to sheltered reef crest/flat zones, backreef zones. Most of the original descriptions of core sedimentology generally lacked sufficient detail to be used directly in this review. In particular, the coral material recovered in the cored sections was identified by the authors, either only on the basis of gross colony shape or at a variety of taxonomic levels, from order to species. Given that a comparison of reef facies successions from different locations requires standardization of core logs, careful re-analysis of the published core data was

conducted to ensure consistency in the description of community composition and of facies type. Core selection was based on the following criteria: (1) the lithology, nature of components and the dominant growth forms of corals have clearly been defined by the authors; (2) in a given section, radiometric dates numerically are sufficient for accretion history to be reconstructed; (3) for each drilling site, patterns of modern coral community composition and distribution are accessible from the literature. Identification of the fossil coral assemblages from core descriptions was attempted on the basis of coral growth shapes and by comparison with present-day coral zonations and ecologies. With this aim, various regional and general monographs on the taxonomy and/or the distributional patterns of living corals were compiled: Western Indian Ocean (Rosen, 1971a,b; Pichon, 1978; Faure, 1982), Central and Eastern Indian Ocean (Scheer, 1971; Pillai, 1971; Pillai and Scheer, 1976), Japan (Takahashi et al., 1985; Nakamori, 1986; Veron, 1992; Iryu et al., 1995), Eastern Australia (Done, 1982; Veron, 1986), New Caledonia (Chevalier, 1971, 1975), Central Pacific (Wells, 1954; Chevalier, 1979; Faure and Laboute, 1984; Maragos and Jokiel, 1986; Maragos, 1974), Eastern Pacific (Glynn and Ault, 2000), Indo-Pacific (Veron, 1993, 2000; Wallace, 1999). The coral assemblages so recognized were named after characteristic dominant forms. The cores presented in Figs. 2–4 are considered to provide clear pictures of the internal structure and growth histories of the reef site from which they were extracted. Given the variation in the quality of the core record and descriptions, it would be unrealistic to use these data to make definitive statements on the history of Indo-Pacific reef development.

In other respects, in order to investigate the nature and topography of the pre-Holocene foundations beneath modern reefs or the stratal patterns within the Holocene sedimentary sequences, drilling investigations have locally been combined with seismic surveying. Unfortunately, seismic stratigraphy has been used in a limited number of Indo-Pacific sites (Mayotte: Zinke et al., 2001; Central Red Sea: Dullo and Montaggioni, 1998; Maldives: Purdy and Bertram, 1993; Great Barrier Reef: Harvey, 1986; Tasman Sea: Woodroffe et al., 2004; Mururoa: Guille et al., 1996).

Table 2  
 Depths of occurrence, morphology and ages of positively identified Postglacial submerged reefs in the Indo-Pacific province

Location	Present depth (in m)	Height (in m)	Radiometric ages (in 1000 cal. years BP)	Reef type	Expected ages <sup>a</sup> (1000 years BP)	Source
Mayotte <sup>b</sup> , Comoro islands	55–65	NA	NA	mounds		Colonna et al. (1996), Dullo et al. (1998)
	90	>3	10.10	mounds		
	100–112	>3	13.60	mounds		
	118	NA	16.90–18.20			
	152	NA	18.40			
Western Indian shelf	70	NA	8.82–9.82	NA		Vora et al. (1996), Rao et al. (2003)
	85–100	8–15	12.75–12.89	barriers, pinnacles		
	105–110	1–6	13.68–15.41	barriers, pinnacles		
	120–136	2–6	NA	pinnacles		
Eastern Australian shelf	25–30	2–15		pinnacles	9–10	Carter and Johnson (1986), Harris and Davies (1989), Hopley et al. (1997)
	40–45	<10		barriers, pinnacles	10–11	
	50–64	2–15	modern–0.66 <sup>c</sup>	barriers, pinnacles	12	
	70–75	2–35		barriers, pinnacles	15	
	90–100	<5		pinnacles	17	
	120–130	NA		pinnacles	18	
	175	NA	13.60–17.00	pinnacles		
Hawaiian islands Maui, Lanai	18	>10		pinnacles	8	Grigg et al. (2002)
	37	4–12		pinnacles	10	
	63	NA		pinnacles	12	
	82	NA			14	
	120	NA			21	
Hawaii <sup>d</sup>	150–160	NA	13.90	NA		Moore et al. (1990)
	207	NA	15.80	NA		
Molokai	100–150	>15–>30	14.70–15.20	barrier reef		Webster et al. (2004) Engels et al. (2004)
	17.7	>1.50	>7.96	fringing reef		
	21	>1	>8.14	fringing reef		Engels et al. (2004)
French Polynesia						
Marquesas	90–100	NA	20.00	fringing reef		Cabioch et al. (2000, 2003b)
Mururoa	167	NA	17.50	NA		Camoïn et al. (2001)

<sup>a</sup> The expected ages are speculative, based on estimates of paleosea-level position during the Postglacial sea level rise (from sea level curves by Fairbanks, 1989; Edwards et al., 1993; Bard et al., 1996).

<sup>b</sup> Subsiding island at a rate of 0.25 mm year<sup>-1</sup>.

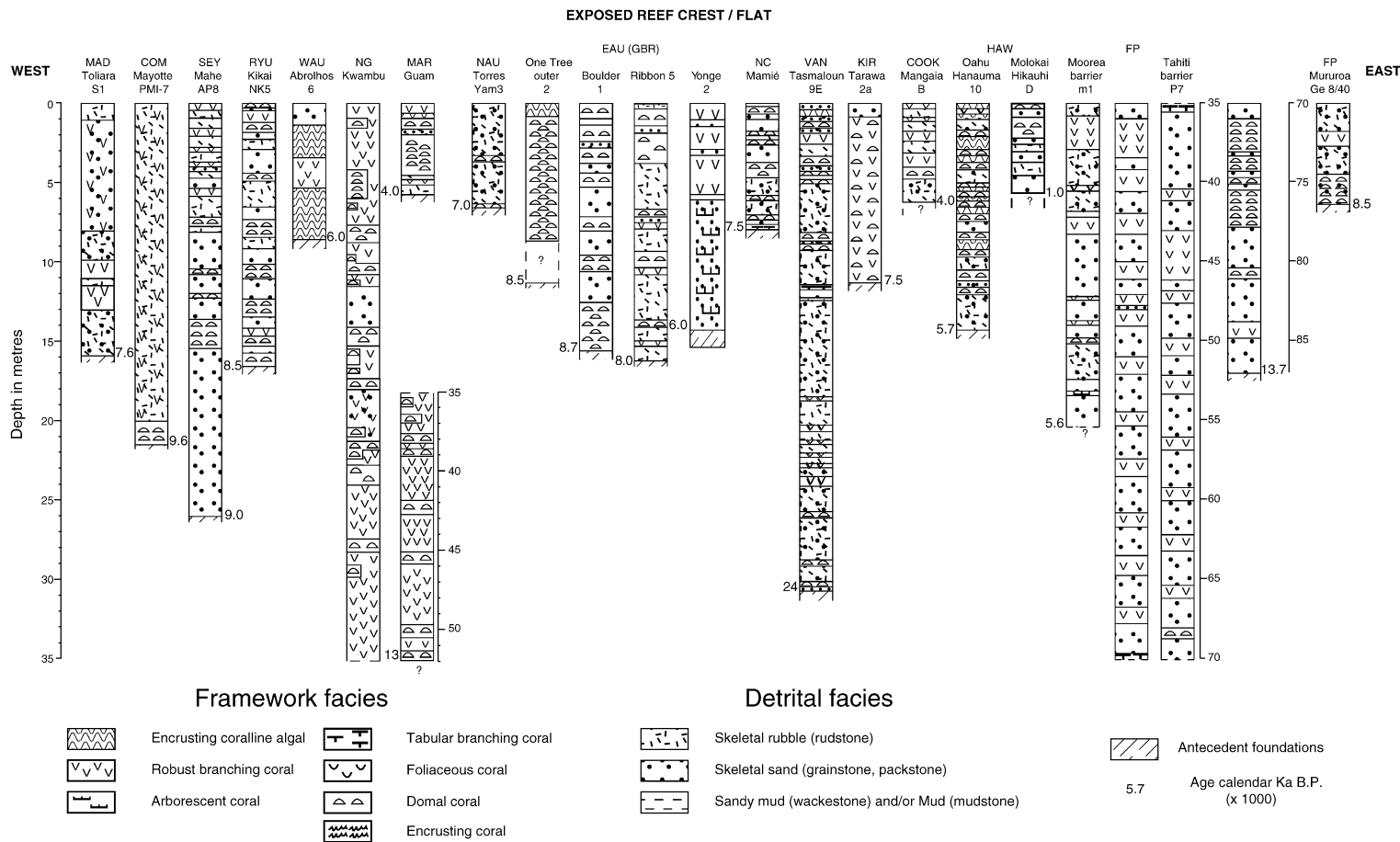
<sup>c</sup> Dates obtained from rubble on submerged reefs.

<sup>d</sup> Subsiding island at a rate of 2.5 mm year<sup>-1</sup>.

## 2.2. Accuracy of radiometric dates

Recent to Late Pleistocene corals can be dated by radiocarbon and uranium-series disequilibrium methods (alpha spectrometry and thermal ionization mass spectrometry—TIMS). Samples that have remained bathed in seawater up until the time of drilling generally give consistent results (Quinn and Tudhope, 2000). The chronology of most of the Postglacial reef

sections has been mainly determined by radiocarbon dating. Less than 80 cores out of the 684 compiled were dated using U-series analyses. Prior to the 1980s, <sup>14</sup>C ages reported in the literature corresponded to conventional ages calculated with the conventional 5568-year half-life and were uncorrected for the metabolic fractionation of <sup>12</sup>C/<sup>13</sup>C ratios and for secular atmospheric variations in <sup>14</sup>C. To consistently compare age datasets based on radiocarbon and U-



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Fig. 2. Bio-lithologies of selected shallow-water cores from exposed reef crest/flat zones in the Indo-Pacific reef province. Each core is identified by the hole number, the name of the reef site and the relevant geographical area given in the literature (MAD=Madagascar; COM=Comoro islands; SEY=Seychelles; RYU=Ryukyu islands; WAU=Western Australia; NG=New Guinea; MAR=Mariana islands; NAU=Northern Australia; EAU (GBR)=Eastern Australia (Great Barrier Reef); NC=New Caledonia; VAN=Vanuatu islands; KIR=Kiribati islands; COOK=Cook islands; HAW=Hawaii islands; FP=French Polynesia). The data are extracted from Montaggioni et al. (1997b), Camoin et al. (2004) (Toliana, Madagascar), Camoin et al. (1997) (Mayotte, Comoros), Braithwaite et al. (2000) (Seychelles), Webster et al. (1998) (Kikai-Jima island, Ryukyus, Japan), Collins et al. (1993) (Abrolhos, Southwest Australia), Chappell and Polach (1991) (Kambu, Huon Peninsula, New Guinea), Kayanne et al. (1993) (Guam, Mariana Islands), Woodroffe et al. (2000) (Torres Strait, Northern Australia), Davies and Hopley (1983) (Ribbon 5 and Boulder reefs, Central Great Barrier Reef), Webster and Davies (2003) (Ribbon 5), Marshall and Davies (1982) (One Tree Reef, Southern Great Barrier Reef), Laurenti (1995) (Yonge Reef), Cabioch et al. (1995) (Mamié-Ounia reef, New Caledonia), Cabioch et al. (1998) (Tasmaloun, Spiritu Santo, Vanuatu), Marshall and Jacobsen (1985) (Tarawa island, Kiribati), Yonekura et al. (1988) (Mangaia, Cook islands), Easton and Olson (1976) (Hanauma reef, Oahu, Hawaii), Engels et al. (2004) (Molokai island, Hawaii), Montaggioni (1988) (Moorea, Society islands), Montaggioni et al. (1997a) (Tahiti, Society islands), E布伦 (1996) (Mururoa atoll, Tuamotus). The number at the base of each core log refers to the time of reef initiation in the considered site (ages in calendar years×1000 BP) (ka).

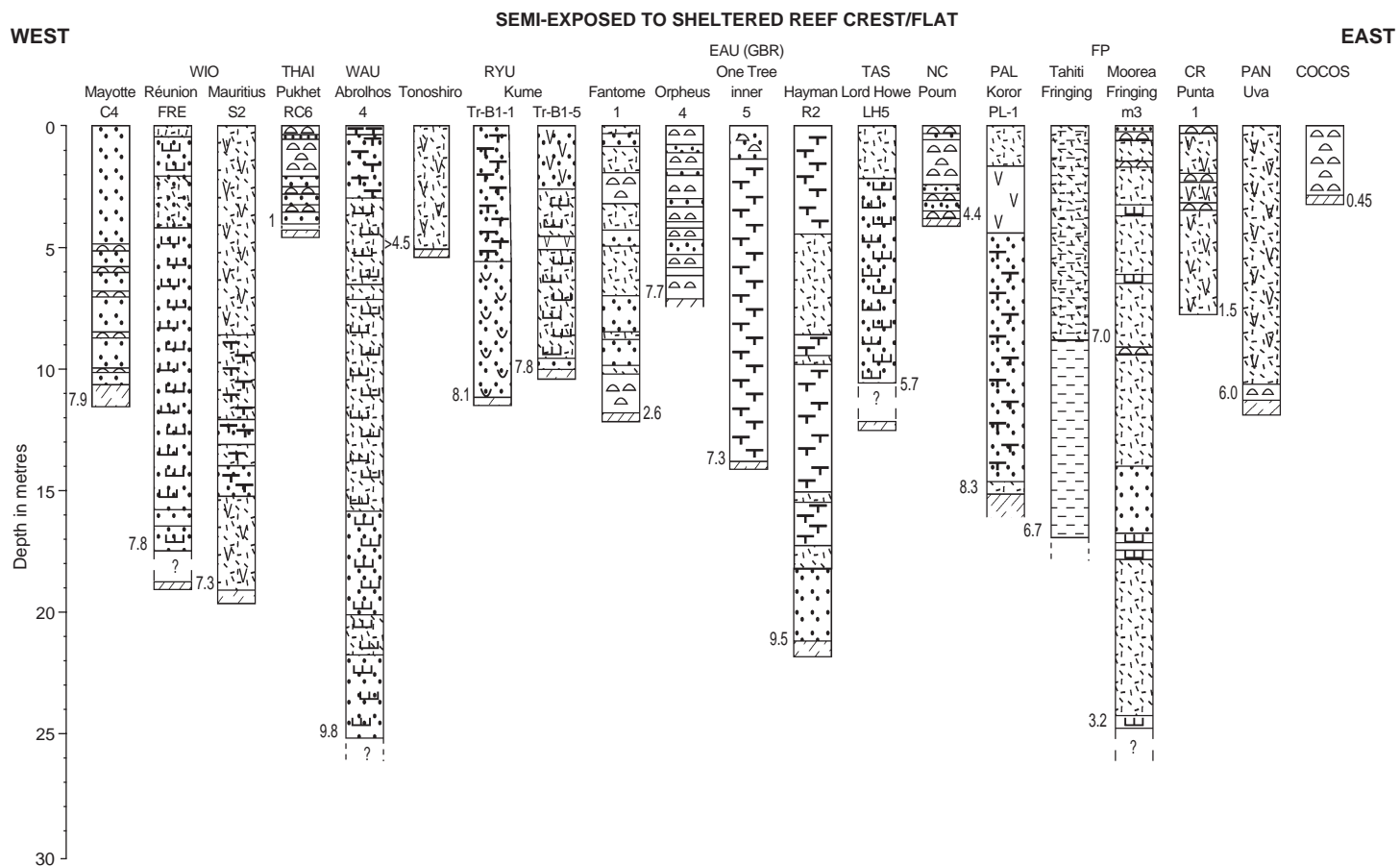


Fig. 3. Bio-lithologs of selected shallow-water cores from semi-exposed to sheltered reef crest/flat zones in the Indo-Pacific reef province. Each core is identified by the hole number, the reef site and the relevant geographical area given in the literature. WIO=Western Indian Ocean; THAI=Thailand; WAU=Western Australia; RYU=Ryukyu islands; EAU (GBR)=Eastern Australia (Great Barrier Reef); TAS=Tasman Sea; NC=New Caledonia; PAL=Palau islands; FP=French Polynesia; CR=Costa Rica; PAN=Panama; COCOS=Cocos islands. The data are extracted from Zinke et al., (2003a,b) (Mayotte island), Montaggioni (1977) (Réunion island), Montaggioni and Faure (1997) (Mauritius island), Tudhope and Scoffin (1994) (Pukhet, South Thailand), Collins et al. (1993) (Abrolhos, Southwest Australia), Yamano et al. (2003) (Tonoshiro, Ishigaki island, Ryukyus), Takahashi et al. (1988) (Kume Island, Ryukyus, Japan), Hopley et al. (1978) (Hayman Island, Central Great Barrier Reef), Hopley et al. (1983) (Orpheus island, Central Great Barrier Reef), Johnson and Risk (1987) (Fantome island, Central Great Barrier Reef), Marshall and Davies (1982) (One Tree Reef, Southern Great Barrier Reef), Kennedy and Woodroffe (2000) (Lord Howe island, Tasman Sea), Cabioch et al. (1995) (Poum reef, New Caledonia), Kayanne et al. (2002) (Koror islet, Palau islands), Hein et al. (1997) (Aitutaki, Cook islands), Montaggioni (1988) (Tahiti and Moorea, Society islands), Cortés et al. (1994) (Punta islotes, Costa Rica), Glynn and Macintyre (1987) (Uva island, Panama), Macintyre et al. (1992) (Cocos island, Eastern Pacific). The number at the base of each core log refers to the time of reef initiation in the considered site (ages in calendar ka). See Fig. 2 for symbols.

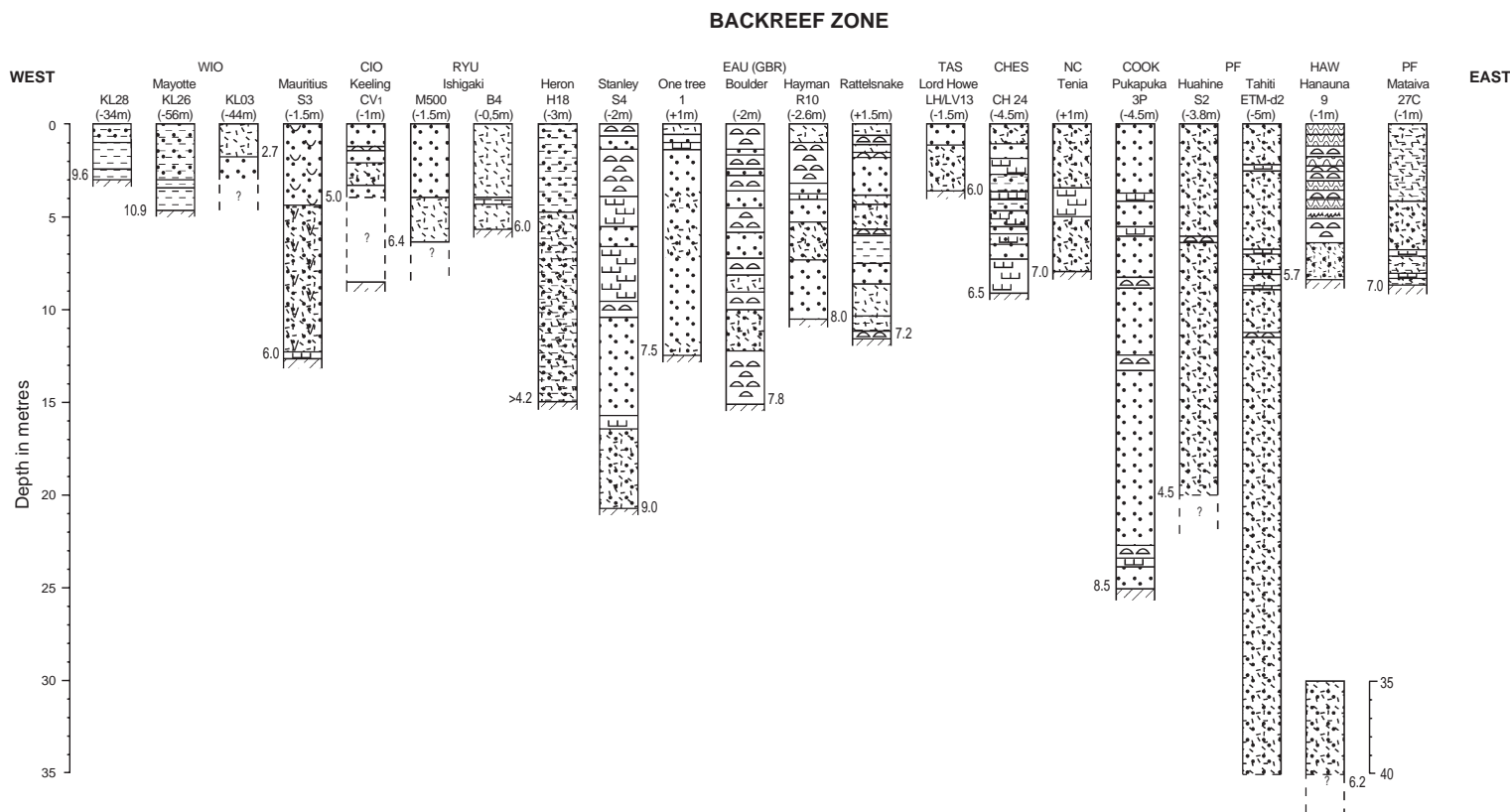


Fig. 4. Bio-lithologies of selected shallow-water cores from backreef zones in the Indo-Pacific reef province. Each core is identified by the hole number, the reef site and the relevant geographical area. WIO=Western Indian Ocean; CIO=Central Indian Ocean; RYU=Ryukyu islands; EAU (GBR)=Eastern Australia (Great Barrier Reef); TAS=Tasman Sea; CHES=Chesterfield islands; NC=New Caledonia; COOK=Cook islands; PF=French Polynesia; HAW=Hawaiian islands. The data are extracted from Zinke et al. (2003a,b) (Mayotte island), Montaggioni and Faure (1997) (Mauritius island), Woodroffe et al. (1999) (Cocos-Keeling islands), Yamano et al. (2001a) (Ishigaki Island, Ryukyus, Japan), Smith et al. (1998) (Heron island, Southern Great Barrier Reef), Davies et al. (1985) (Stanley reef, Central Great Barrier Reef), Marshall and Davies (1982) (one tree Reef, Central Great Barrier Reef), Webster (1999) (Boulder reef, Central Great Barrier Reef), Hopley et al. (1978) (Hayman and Rattelsnake islands, Central Great Barrier Reef), Kennedy and Woodroffe (2000) (Lord Howe island, Tasman Sea), Degauge-Michalski (1993) (Chesterfield islands), Coudray (1976) (Tenia islet, New Caledonia), Gray et al. (1992) (Pukapuka atoll, Cook islands), Montaggioni, unpublished (Huanine, Society islands), Montaggioni (1988) (Tahiti, Society islands), Easton and Olson (1976) (Oahu island, Hawaii), Pirazzoli and Montaggioni (1986) (Mataiva atoll, Tuamotus). The number at the base of each core log indicates the time of reef initiation in the considered site (ages in calendar ka). The values (in m) within brackets refer to water depth at each core site. See Fig. 2 for symbols.

series methods, respectively, all of the  $^{14}\text{C}$  dates used in this review (Tables 1 and 2) were reconverted into calibrated calendar years (cal. years BP) using the CALIB-4 software programme (Stuiver et al., 1998). This calibration protocol assumes that the mean marine global  $^{14}\text{C}$  reservoir age has remained constant through time (ocean reservoir value: 402 years). This assumption has recently been questioned (Sikes et al., 2000); compared to modern values, surface-reservoir ages at 11.9 ka radiocarbon age BP, ranged around 800 years and during the LGM, reached 2000 years. Calendar calibration at present extend through the Holocene to 20.4  $^{14}\text{C}$ -years BP (equivalent to 24 cal. years BP in CALIB 4) (Mix et al., 2001). The calibrated ages are 0 to 500 years older than the conventional radiocarbon ages within the 4–8 cal. ka BP range, and about 1 to 3.5 years older within the 9–24 cal. ka BP interval. Although the accuracy of the older radiocarbon data remains uncertain, the calibrated ages can be directly compared with those determined by U-series methods, since the U–Th dates require no reservoir or calendar corrections. In addition, U–Th TIMS dating provides high-precision values that range from 2 years in a 100-year-old sample to less than 100 years in a 23,000-year-old sample.

### 3. Reef development patterns

#### 3.1. Reef facies

The analysis of the core dataset allowed 10 major facies to be delineated, including 7 varieties of growth framework and 3 detrital facies. These facies are named after the nature and the growth shapes of the dominant builders. Their interpretation is based on comparisons of the biolithologic components of the cores with those found in the modern reef zones (Fig. 5A,B).

##### 3.1.1. Coralline algal facies

This facies is largely restricted to cores coming from windward reef crests and reef flats, and from lagoonal patches that just face passes trenching reef rims. In these zones, calcareous algae form centimetre- to decimetre-thick laminated crusts building a dense and homogenous framework, or thinner crusts

binding in-place or reworked corals. These crusts can reach 1 to up to 2 m in total thickness (Easton and Olson, 1976; Marshall and Davies, 1982; Marshall and Jacobsen, 1985; Takahashi et al., 1985; Kleypas and Hopley, 1993; Collins et al., 1993; Macintyre, 1997; Montaggioni et al., 1997a; Grigg, 1998; Kayanne et al., 2002; Yamano et al., 2003; Grossman and Fletcher, 2004). In cores extracted from more sheltered areas (i.e. inner reef flats, backreef patch reefs), coralline algae have generally developed as branching veneers on corals. Irrespective of their thickness, the coralline crusts have mainly been formed by *Hydrolithon onkodes*, *Neogoniolithon* spp., *Mesophyllum* sp. and *Lithophyllum* sp., associated with encrusting foraminifers, vermetid gastropods and serpulids. During the rise of sea level, conditions favourable for the growth of thick coralline crusts operated mainly during “keep-up” stages of reef accretion and particularly in the final growth phase after sea-level stabilized. During “catch-up” growth phases, corallines have grown in the form of millimetre-thick crusts over corals and skeletal debris, irrespective of the environmental setting (Marshall and Davies, 1982; Macintyre, 1997; Cabioch et al., 1999a,b; Yamano et al., 2003). However, in this case, because of the low core volume occupied (less than 5% of the recovered material), these cannot be interpreted as a differentiated facies.

##### 3.1.2. Robust-branching coral facies

Previously referred to as branching, stout-branching, stubby-branching, and even coral limestone, this facies originally forms an open to cavernous interlocking framework of thick-branched, wave-resistant corals dominated by acroporids (*A. robusta* group; *Acropora palifera*; *Acropora humilis* group) and pocilloporids (*P. damicornis*, *P. verrucosa*, *Stylophora pistillata*), with associated domal poritids (*Porites* spp.), faviids (*Goniastrea*, *Favia*, *Platygyra*) and siderastreids (*Psammocora stellata*). The corals in this assemblage commonly are thickly encrusted by crustose corallines (mainly *Hydrolithon* cf. *onkodes*) and arborescent foraminifers (Homotremids, Victoriellids and/or Acervulinids). The framework is usually extensively bored by molluscs (vermetids and sponges). Cavities up to 0.10 m high are widespread within the framework. The larger ones are filled mainly by rubble derived from acroporid and pocillo-



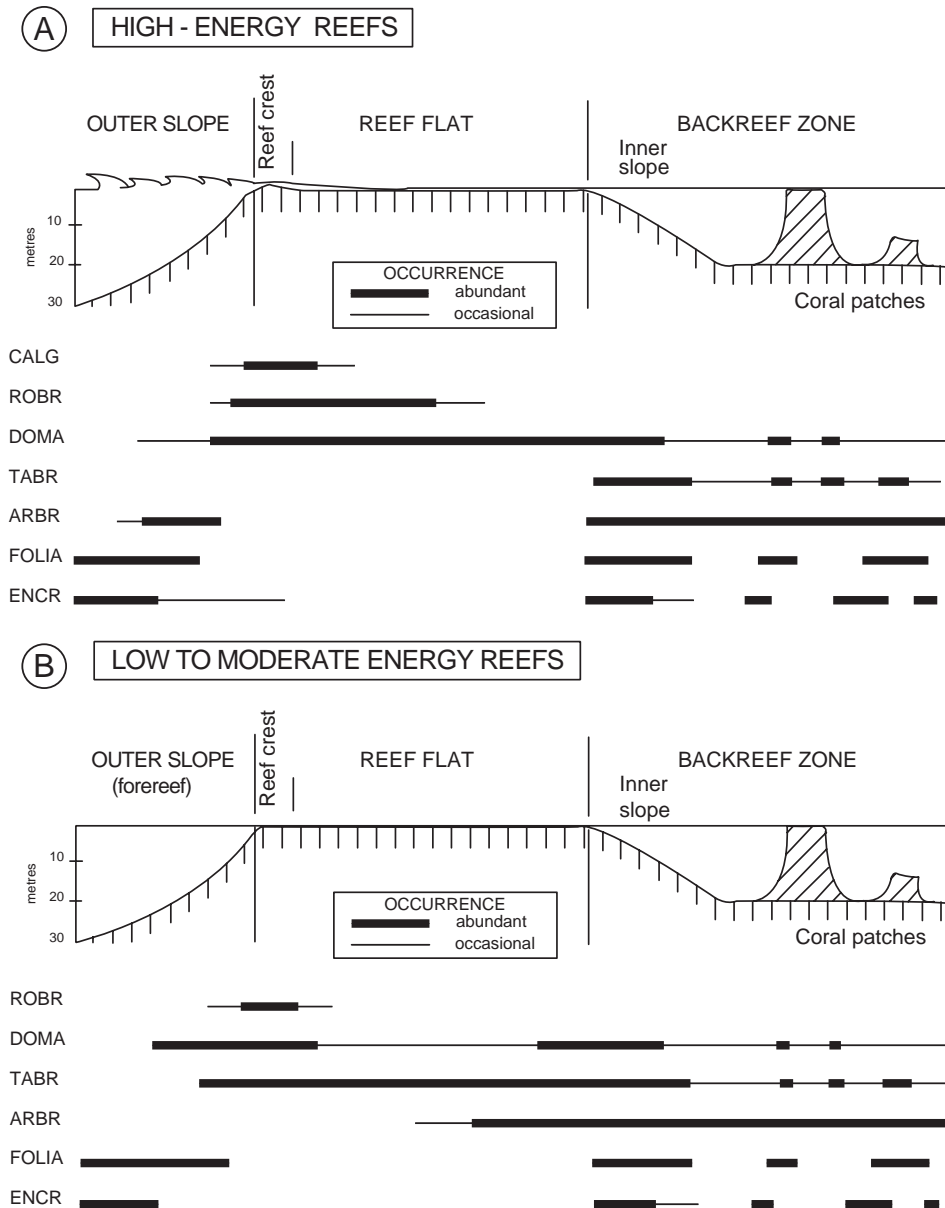


Fig. 5. Schematic summary of distribution of the dominant reef-building coral (and coralline algal) forms and assemblages in the Indo-Pacific tropics, from deeper outer slopes to backreef zones in relation to water energy and water depth. (A) Distribution in reef systems subject to higher hydrodynamic energy (outer shelf settings). (B) Distribution in reef systems subject to lower hydrodynamic energy (protected, open and mid- to inner-shelf settings). Assemblages and facies : CALG=coralline algal; ROBR=robust-branching coral; DOMA=domal coral; TABR=tabular-branching; ARBR=arborescent coral; FOLIA=foliaceous coral; ENCR=encrusting coral.

porid colonies in a sandy matrix. The latter is composed generally of tests of benthic free-living foraminifers (*Amphistegina*, *Calcarina* and/or *Baculogypsina*), encrusting foraminifers (*Homotrema*,

*Carpenteria*), green algal *Halimeda* plates and fragments of corals, molluscs and corallines. Lithified micritic crusts within framework cavities were observed in various Indo-Pacific sites (Camoin et al.,

1999; Marshall and Davies, 1982); they have markedly reduced intraskeletal porosity. Successive boring events, repeated sediment infilling and cementation have produced a heterogeneous rock.

There is little geographic variations in this facies. In cores from the Western Indian Ocean, the dominant builders belong to the groups of *A. robusta* and *A. humilis* (Cabioch et al., 1999b; Camoin et al., 2004). In the Western Pacific, *A. palifera* locally has been one of the main framework components, particularly in the Northern Great Barrier Reef of Australia (Webster and Davies, 2003), in New Caledonia (Cabioch et al., 1995) and in Papua New Guinea (Ota and Chappell, 1999). In the Ryukyus, *Acropora gemmifera* and *Pocillopora verrucosa* are the most abundant species (Sugihara et al., 2003). In Western Pacific oceanic islands, the dominant framework producers have been *A. gr. humilis* (Kayanne et al., 2002). The framework in Tahiti island principally comprises *A. gr. robusta*, associated with *P. damicornis* (Pirazzoli and Montaggioni, 1988; Montaggioni and Camoin, 1993; Cabioch et al., 1999a,b) and in Mururoa atoll, Tuamotus as well (Perrin, 1989; Camoin et al., 2001). By contrast, in Hawaii islands, the acroporid forms seem not to have contributed to framework growth; the dominant robust-branching corals are represented by *Pocillopora* sp. (Easton and Olson, 1976), *P. meandrina* (Grigg, 1998) and *Porites compressa* (Engels et al., 2004). A similar scheme was found in the Eastern Pacific; Glynn and Macintyre (1987) and Cortés et al. (1994) pointed out that the robust-branching coral facies from Panama and Costa Rica reefs has been made up of *P. damicornis*.

This facies is the fossil counterpart of the coral community including dominant acroporids (*A. gr. robusta*: *A. robusta*, *A. danai*, *A. abrotanoides*, *A. palmerae*; *Acropora* (*Isopora*) *palifera*; *A. gr. humilis*: *A. humilis*, *A. digitifera*, *A. gemmifera*; *Acropora latistella*) and pocilloporids (*P. damicornis*, *P. eydouxii*, *P. verrucosa*, *P. meandrina*, *S. pistillata*, *S. mordax*). Subordinate associated corals are domal (*Porites lutea*, *P. lobata*, *Leptoria phrygia*, *Platygyra daedala*, *Goniastrea retiformis*, *Goniastrea favulus*, *Favia stelligera*, *Favia pallida*, *Favia flexuosa*, *Psammocora* sp., *Astreopora* sp., *Montipora* sp.), platy (*Millepora platyphylla*), tabular (*A. hyacinthus*), columnar (*Porites annae*) and encrusting (*Montipora tuberculosa*, *Echinopora gemmacea*).

This community is encountered as a distinct zone on modern windward reef crests, proximal reef fronts and, less commonly, reef flat environments in water depths of less than 6 m in most of the Indo-Pacific areas: Red Sea (Riegl and Piller, 2000), Western Indian Ocean (Rosen, 1971a,b; Scheer, 1971; Pichon, 1978; Faure, 1982; Montaggioni and Faure, 1980, 1997), Eastern Australia (Done, 1982; Veron, 1986; Van Woesik and Done, 1997), Ryukyu islands (Nakamori, 1986; Iryu et al., 1995), Palau islands (Kayanne et al., 2002), Papua New Guinea (Nakamori et al., 1995), Central Pacific (Chevalier, 1979; Maragos, 1977; Bouchon, 1985; Faure and Laboute, 1984; Faure, 1986; Maragos and Jokiel, 1986; Cabioch et al., 1999a). The diversity of the robust-branching coral assemblage drastically decreases eastward (Glynn and Macintyre, 1987; Cortés et al., 1994; Glynn and Ault, 2000). Thus, the paleoenvironment of the robust-branching coral facies is regarded as high-energy, very shallow (0–6 m deep), diagnostic of windward margins (upper forereef to outer reef flat zones).

### 3.1.3. Domal coral facies

Previously known as coral head or massive, this facies has been identified in cores on the basis of the dominance of dome-shaped poritids (*P. lutea*, *P. lobata*, *Porites* sp., *Goniopora* spp., *Cyphastrea* spp.), faviids (*Favia speciosa*, *F. stelligera*, *Favia* spp., *G. retiformis*, *Goniastrea* spp., *Platygyra sinensis*, *P. daedala*, *Diploastrea heliopora*, *Favites* spp.), acroporids (*Astreopora listeri*) and/or mussids (*Symphillia*). The subordinate forms can be branching *Galaxea fascicularis* and encrusting *Montipora capitata*, *Montipora* sp. These corals are locally mixed with debris of various acroporids. Two subfacies can be differentiated from the degree of development of associated coralline algal crusts. A robust encrusted subfacies with coralline crusts 1–10 cm thick is found in cores taken from windward reef flats (Easton and Olson, 1976; Marshall and Davies, 1982; Yonekura et al., 1998; Cabioch et al., 1995, 1999a,b; Cabioch, 2003). A thin- or non-encrusted subfacies in which coralline crusts are about 1 mm thick or absent appears in cores extracted from sheltered reef flats, patch reefs and backreef zones, peculiarly those subjected to turbid waters (Hopley et al., 1983; Buigues, 1985; Johnson and Risk, 1987; Tudhope and Scoffin, 1994; Kleypas, 1996; Montag-

gioni and Faure, 1997; Braithwaite et al., 2000; Woodroffe et al., 2000; Kayanne et al., 2002; Zinke et al., 2003a,b; Camoin et al., 2004; Grossman and Fletcher, 2004). Thus, the facies relates either to a dense framestone composed of coral heads separated by composite crusts of corallines with vermetids, serpulids and arborescent foraminifers or to in-situ, scattered massive colonies entombed in a porous matrix of skeletal sand and silt. Corals usually are bored by bivalves, sponges and polychaetes. The matrix is composed of a variety of bioclasts among which alcyonarian spicules, large benthic foraminifers (amphisteginids, calcarinids) and *Halimeda* plates predominate. Cemented mud is present occasionally within large voids.

This coral assemblage displays a low regional variability throughout the Indo-Pacific province, except in environments subjected to extreme conditions (low temperature, high mud input). In this case, the coral fauna is severely depauperate; a few species only have participated in reef building (Tudhope and Scoffin, 1994; Yamano et al., 2001a,b). In the far eastern Pacific, the domal coral facies has resulted from the growth of the species *Porites lobata* or *Pavona gigantea* (Macintyre et al., 1992; Cortés et al., 1994).

In the Indo-Pacific modern reefs, the distribution of coral communities dominated by massive poritids and faviids is widespread. The latter occur within the 0–25-m depth range, on semi-exposed to sheltered, windward to leeward reef slopes and reef flats in both outer and inner shelf settings and in backreef slopes and bottoms. The dominant corals may include the poritids *P. lutea*, *P. lobata*, *P. cylindrica*, the faviids *Favia fava*, *F. stelligera*, *F. speciosa*, *Favites abdita*, *Cyphastrea* spp., *Goniastrea pectinata*, *G. edwardsi*, *D. heliopora*, *Montastrea curta*, the acroporid *A. listeri*, with associated mussids (*Symphillia recta*), merulinids (*Hydnophora microconos*) and acroporids (*A. listeri*, *Acropora* spp.) (Rosen, 1971a,b; Scheer, 1971; Chevalier, 1979; Done, 1982; Faure, 1982; Bouchon, 1985; Veron, 1986; Montaggioni and Faure, 1997; Cabioch et al., 1999b; Riegl and Piller, 2000). Locally these occur with a variety of other corals reflecting differing water agitation. In shallow, higher wave-energy areas, the communities are composed largely of *P. lobata* (Grigg, 1998) or augmented by robust-

branching (*A. gr. robusta* and *humilis*, *A. palifera*, *A. bruggemanni*, *S. pistillata*). The relevant corals are usually bound postmortem by thick coralline crusts.

In less agitated or deeper waters, coral forms include domal (*P. lobata*, *Pseudosiderastrea tayamai*, *Moseleya latistellata*), tabular (*A. gr. hyacinthus*), delicate branching (*Acropora divaricata*, *A. splendida*, *A. muricata*, *Seriatopora hystrix*) and/or foliaceous (*M. capitata*, *M. aequituberculata*), laminar (*Montipora verrucosa*) and columnar colonies (*Porites nigrescens*). Encrusting coralline algae are poorly developed.

Although the present-day poritid/faviid communities are widely distributed, they tend to occupy lower energy reef settings, from surface to depths of around 10–15 m (Rosen, 1971b; Done, 1982; Faure, 1982; Cabioch et al., 1999b; Grossman and Fletcher, 2004). This strongly suggests that the domal coral facies has formed in a semi-exposed to sheltered, relatively deeper paleoenvironment.

#### 3.1.4. Tabular-branching coral facies

Previously referred to as table, platy, tabulate coral, this facies was encountered in cores either as a loose open or a rigid structural framework, entombed in a porous matrix of gravels, sands and silts. The corals include plate-shaped acroporids (*A. gr. hyacinthus*), associated with a variety of other branching forms (*A. gr. humilis*, *Acropora* spp., *P. verrucosa*, *Pocillopora* spp.), massive or digitate poritids, and domal faviids. According to the state of preservation of the original coral assemblages, this facies can be subdivided into two subfacies: (1) an in-situ tabular subfacies, identified owing to the presence of pieces of colonies the concave surfaces of which are orientated upwards; these corals are weakly affected by boring and abrasion, but separated by lithified bioclasts. Locally coralline algae form crusts less than 5 mm thick over coral elements. (2) A reworked tabular subfacies, exhibiting rubble of platy acroporids, abraded and intensively bored that are mixed with unconsolidated or firmly cemented silty sands. In both subfacies, the matrix is poorly sorted and may be composed of coral, molluscan, echinoid and *Halimeda* grains, with free-living foraminiferal tests (amphisteginids, calcarinids, baculogypsinsids) and debris of branched corallines.

The tabular-branching facies is widely distributed all the way from the western Indian Ocean to the Central Pacific. It pre-eminently constitutes the framework of reef-front, reef-flat and reef-patch structures in windward to leeward, low to moderate wave-energy settings (Hopley et al., 1983; Marshall and Davies, 1982; Yonekura et al., 1988; Collins et al., 1993, 2003; Montaggioni and Faure, 1997; Cabioch et al., 1995, 1999a,b; Yamano et al., 2001a; Kayanne et al., 2002; Cabioch, 2003). There is no significant change in the facies composition from site to site. The *A. hyacinthus* group remains the major facies contributor. By contrast, this facies has not been encountered in the Eastern Pacific reefs (Glynn and Macintyre, 1987; Macintyre et al., 1992); this is certainly due to the absence of coral species giving tabular growth forms in this region (Glynn and Wellington, 1983; Glynn and Ault, 2000). While species of the *A. hyacinthus* group were reported from modern reefs surrounding these islands (Wallace, 1999), the Holocene reef sequences drilled in the Hawaiian islands (Easton and Olson, 1976) and the southeastern Tuamotus (Perrin, 1989; Camoin et al., 2001) did not reveal the occurrence of tabular corals, probably because of the scarcity of the relevant species.

The tabular-branching coral facies is the analog of the present-day acroporid-dominated community that consists of *A. gr. hyacinthus* (*A. hyacinthus*, *A. cytherea*, *A. subulata*), associated with a variety of other acroporids (*A. splendida*, *A. intermedia*, *A. humilis*, *A. digitifera*, *A. nobilis*, *A. squarrosa*, *Montipora digitata*, *S. pistillata*), pocilloporids (*P. verrucosa*, *P. damicornis*, *P. eydouxi*) and poritids (*P. nigrescens*, *P. lutea*). Subordinate forms include domal *Lepastrea* and *Platygyra*, columnar *Alveopora* and laminar *Echinophyllia* and *Echinopora*. This community occurs in semi-exposed or sheltered, upper and mid-forereef zones, reef flats and adjacent backreef slopes and patches, usually in mid-shelf situations. It is growing intertidally or subtidally at depths not exceeding 20 m; however, in most cases, this coral assemblage occupies the 2–15-m depth interval (Scheer, 1971; Pichon, 1978; Done, 1982; Faure, 1982; Faure and Laboute, 1984). Thus, in the fossil record, the tabular-branching coral facies is thought to represent an open marine, moderate wave-energy environment at depths not exceeding 15 m (Cabioch et al., 1999b).

### 3.1.5. Arborescent coral facies

Also previously called “staghorn-like”, ramose, thin-branched, delicate or gracile branching, this facies appears in cores either as a framework of upright, thin branched (less than 20 mm in diameter), locally coralline-encrusted or as fresh angular fragments and portions of branches. In this case, fragmentation and reworking of the relevant colonies can be clearly recognized as resulting from rotational drilling. Coralline algae play the same role as in the tabular-branching coral facies; they have developed in the form of millimetre-thick coats over coral clasts, in association with sessile foraminifers. The dominant scleractinians are *A. gr. muricata*, *Acropora gr. aspera*, *A. gr. divaricata* and *P. compressa*. Other contributors include domal and digitate poritids (*P. nigrescens*), a variety of branching forms (*A. gr. hyacinthus*, *Pocillopora* spp.) and a few foliaceous species. The coral material recovered may be loose or partly lithified; it is mixed with a porous matrix of unconsolidated to poorly cemented bioclastic sands and silts. These grains are derived mainly from corals, molluscs, articulated corallines (*Amphiroa*, *Coralina*), *Halimeda* particles and foraminiferal tests (miliolids, soritids, calcarinids). The arborescent coral facies was found in drill holes penetrating modern inner reef flats and adjacent backreef slopes, in both outer- and mid-shelf settings and reef-front slopes in inner-shelf areas.

This facies shows little variations throughout the province. In the Western Indian Ocean, the framework components are *A. gr. muricata*, *A. gr. aspera*, associated with *A. gr. hyacinthus* (Montaggioni, 1977; Camoin et al., 1997). Similar assemblages were identified in the Holocene sections from different sites: Ryukyus (Yamano et al., 2001a,b, submitted for publication); Palau (Kayanne et al., 2002); inner-shelf reefs, Northeast Australia (Hopley and Barnes, 1985; Kleypas, 1996), Lord Howe (Kennedy and Woodroffe, 2000), Chesterfield Islands (Degauge-Michalski, 1993), Society Islands (Montaggioni, 1988), Tuamotus (Pirazzoli and Montaggioni, 1986; Perrin, 1989). The facies in question seems to be restricted to delicate branching *P. compressa* framestones in the Holocene sequences in the Hawaiian Islands (Grigg, 1998; Grossman and Fletcher, 2004). It is missing from the eastern Pacific sites (Glynn and Macintyre, 1987; Macintyre et al., 1992).



The aborescent coral facies is the counterpart of present-day branching coral communities that house lower to middle parts of fore-reef zones, inner reef flats and nearby backreef slopes in semi-exposed to sheltered environments. Along fore-reef slopes, the arborescent coral community may predominantly include *A. gr. divaricata* (*A. divaricata*, *A. clathrata*), *A. aculeus*, *A. valenciennesi*, *A. tenuis*, *S. hystrix*, and *S. pistillata*. On reef-flat and backreef zones, the coral community may be dominated by large thickets of *A. gr. muricata* (*A. muricata*, *A. grandis*), *A. gr. aspera* (*A. aspera*, *A. pulchra*), *A. cerealis*, *A. valida*, *A. tortuosa*, *A. austera*, *A. intermedia*, *A. microphthalma*, *A. gr. lovelli*, *S. hystrix*, *S. pistillata*, *P. damicornis*, *Echinopora horrida*. In both types of habitat, the subordinate forms may consist of *A. gr. hyacinthus*, *G. pectinata*, *Acropora squarrosa*, *P. lutea* (Scheer, 1971; Chevalier, 1979; Done, 1982; Faure, 1982, 1986; Bouchon, 1985; Delesalle, 1985; Nakamori, 1986; Veron, 1986, 1992; Wallace, 1999). Given there is not significant pattern of differences between the two modern arborescent coral communities, distinction from cored material is quite difficult; thus, the fossil relevant facies is interpreted simply as reflecting reef growth in sheltered settings at paleo-depths between 0 and 20 m.

### 3.1.6. Foliaceous coral facies

Also known as lamellar platy, or tabular coral, this facies is composed mainly of loose bifacial fronds and elements derived from “leaf” *Pavona* (*Pavona cactus*, *P. decussata*, *P. varians*), or *Pachyseris* (*P. rugosa*) colonies, and from flat platy and whorled colonies of *Montipora* (*M. foliosa*) and *Turbinaria* spp. The subordinate constructors may include massive faviids, poritids and *Hydnophora* sp., delicate branching acroporids and a variety of encrusting forms (*Pachyseris*, *Echinopora*). These components can be “grain-supported” or embedded in unconsolidated muddy to fine sands; the sand fraction consists of debris of corals, gastropods, bivalves, branched and articulated corallines, echinoids and free-living foraminifers (Miliolids, calcarinids, discorbids, textulariids, nummulitids).

The preservation state of the coral material varies from site to site. In areas where postmortem burying of colonies has rapidly operated owing to heavy sediment supply, corals may be well preserved from

encrustation and bioerosion as observed in cores from inner self settings (Kennedy and Woodroffe, 2000). In most cores taken from other reef zones, foliaceous coral pieces and clasts are encrusted by corallines and have been affected by severe erosion by microborers (Montaggioni and Faure, 1997).

The foliaceous coral facies is encountered in cores from sheltered habitats, in particular, those subjected to high turbidity conditions (Takahashi et al., 1988; Kleypas, 1996; Montaggioni and Faure, 1997). It shows marked variations in composition according to latitude. In the Western Indian Ocean, the coral assemblages are dominated by agariciids (*Pavona* spp.) and acroporids (*Montipora foliosa*) (Montaggioni and Faure, 1997). Similar associations were reported from the Holocene reef sections in the Ryukyus (Kan et al., 1995; Webster et al., 1998). On the Great Barrier Reef of Australia, the dominant “leaf” builders are dendrophylliids (*Turbinaria* spp.) (Kleypas, 1996). While, in the cores retrieved from Vanuatu reefs, the dominant foliaceous coral is *Pachyseris rugosa* (Cabioch, 2003). The facies has not yet been documented from drilled Holocene sequences in the Central Pacific (Marshall and Jacobsen, 1985; Perrin, 1989; Camoin et al., 2001), in contrast with the wide occurrence of relevant communities in the modern environments (Veron, 1986, 2000). The absence of this facies from the eastern Pacific can be attributed to paleobiogeographical constraints (Glynn and Ault, 2000).

The foliaceous coral facies is the counterpart of modern communities dominated by agariciids, dendrophylliids and some platy acroporids, and inhabiting protected zones usually subjected to suspended sediment loadings (i.e. lower irradiance). Along reef slopes in mid- to inner-shelf settings, the communities are composed chiefly of *Montipora aequituberculata*, *Montipora* spp., *Pachyseris speciosa*, *P. rugosa*, *Turbinaria mesenterina*, *T. reniformis*, *T. frondens*, *Merulina ampliata*, in association with frondose pectiniids (*Pectinia alcornonis*), domal faviids (*Favia* spp., *Favites* spp., *G. pectinata*, *Goniastrea* spp., *Cyphastrea* sp.), poritids (*Porites solida*) and merulinids (*H. microconos*), branching pocilloporids (*S. hystrix*, *S. pistillata*, *P. damicornis*) and acroporids (*Acropora splendida*). They occur at depths from 0 to about 15 m and often occupy the upper parts of the habitats colonized by the arborescent coral commu-

ities (Done, 1982; Faure, 1982; Van Woesik and Done, 1997; Yamano et al., 2001b). On inner reef flats and in shallow backreef zones, the foliaceous coral communities are typified by the abundance of platy *Montipora* colonies (*M. tuberculosa*, *M. verrucosa*, *M. danae*) and *Pavona* (*P. cactus*, *P. decussata*, *P. varians*). The subordinate corals may be massive (*P. lobata*, *P. solida*; *F. pallida*, *F. speciosa*; *F. abdita*; *Plesiastrea versipora*; *Lentastrea purpurea*, *L. transversa*; *Cyphastrea ocellina*, *C. seraila*, *Astreopora myriophthalma*) or branched (*P. verrucosa*, *P. paucistellata*; *Psammocora contigua*; *S. pistillata*; *A. muricata*, *A. valida*) (Scheer, 1971; Maragos, 1974; Pichon, 1978; Faure, 1982, 1986; Delesalle, 1985; Maragos and Jokiel, 1986; Faure and Laboute, 1984; Veron, 2000).

By reference to the zonation of its modern analogs, this facies is thought to have formed in sheltered, outer to backreef environments in less than 15 m of water depth.

### 3.1.7. Encrusting coral facies

Also referred to as lamellar coral, this facies has been encountered in cores taken from reef crests and slopes. Grossman and Fletcher (2004) described bindstones composed of in-situ encrusting *P. lobata*, *Montipora patula*, *M. capitata* from a windward Hawaiian reef crest. Yamano et al. (2001b, 2003) reported the occurrence of such a facies type beneath the outer slopes of inner-shelf fringing reefs in the Ryukyus, subject to turbid waters. From continuous boreholes 300 m in length with seaward inclination of 30 to 45°, Perrin (1989) and Ebren (1996) identified an encrusting coral facies along the deep fore-reef zone at Mururoa atoll. Two subfacies were identified. In sheltered, inner reef settings, the framework comprises encrusting faviids (*Cyphastrea* sp.) and pectiniids (*Echinophyllia* sp.), mixed with other faviids (*F. speciosa*, *Favia* sp.) and merulinids (*Hydnophora* sp.); the coral components are contained in a matrix of carbonate and terrigenous sandy mud. In ocean-facing, deep slopes, the facies consists of cemented pieces of lamellar colonies and coral clasts derived from *Montipora* spp., *Pachyseris* sp., *Cyphastrea* sp., *Goniastrea* sp., *Astreopora* sp., and *Porites lichen*. The subordinate coral species include *P. daedala*, *Pocillopora eydouxi*, *F. speciosa*, *F. stelligera* and *Pseudocolumnastrea maldivensis*

(Perrin, 1989). The principal framework support is made up of encrusting foraminifers (*Acervulina inhaerens*), red algae (*Peyssonnelia*) and, to a lesser extent, serpulid polychaetes; they form coatings centimetre-thick or more over corals.

The resulting rock is extensively bored by sponges and polychaetes. Cavities between individual coral–foraminifer sheets contain carbonate sandy mud, with fine to coarse sand-sized debris or tests of foraminifers (acervulinids, soritids, globigerinids) and fragments of bryozoans, serpulids, corals and sponge spicules.

Given the scarcity of outer-reef drillholes data, no statement can be made about geographic variation in the composition of the encrusting coral subfacies. However, both seem to be absent from the Eastern Pacific reefs (Macintyre et al., 1992).

The encrusting coral facies are the counterparts of coral communities that have been observed in a variety of reef environments subject to low light levels and/or strong water agitation. High-energy reef crests, outer and inner slopes of ocean-facing fringing reefs, mid- to inner-shelf reefs are occupied locally by dominantly crustose corals, at depths from about surface to 10 m. According to the region considered, these may include the acroporids *Montipora monasteriata*, *M. capitata*, *M. undata*, *M. patula*, *M. danae*, the agariciids *Leptoseris mycetoseroides*, *P. speciosa*, the pectiniid *Echinophyllia aspera*, the faviids *Leptastrea purpurea*, *Echinopora lamellosa*, *E. gemmacea*, *Cyphastrea serailia*, *C. microphthalma*, *C. ocellina*, the poritid *Alveopora daedala*, the merulinid *M. ampliata*; these encrusting forms may be mixed with dome-shaped species belonging to faviids (*F. pallida*, *F. speciosa*, *Oulophyllia crispa*), acroporids (*A. myriophthalma*, *A. ocellata*, *Astreopora* spp.), poritids (*P. lutea*, *P. lobata*, *Goniopora lobata*, *G. columna*), mussids (*Lobophyllia corumbosa*, *L. hemprichii*, *Acanthastrea echinata*), with finely branching (*Acropora echinata*, *S. hystrix*) or with meandroid forms (*Plerogyra sinuosa*). Free-living corals (*Fungia* spp., *Halomitra* sp., *Herpolitha* sp.) locally are present (Wells, 1954; Chevalier, 1975; Maragos, 1977; Done, 1982; Faure, 1982, 1986; Veron, 1993; Van Woesik and Done, 1997).

The deep outer shelf-reef slopes, from about 20 deep downwards, are colonized usually by commu-



ities typified by the predominance of *Montipora*, *Pachyseris*, *Leptoseris* and/or *Echinophyllia* (Wells, 1954; Barnes et al., 1971; Maragos, 1974; Chevalier, 1975; Pichon, 1978; Maragos and Jokieli, 1986; Faure, 1982, 1986; Faure and Laboute, 1984; Bouchon, 1985). Along steepwalls, these coral associations can extend upwards to around 8 m of water depth as a result of a drastic decrease in light supply. The dominant species in the communities may be *M. aequituberculata*, *M. verrucosa*, *P. speciosa*, *Leptoseris incrustans*, *L. hawaiiensis*, *L. scabra*, *L. mycetoseroides*, *E. aspera*, *E. echinata*, *Oulophyllia crispa*, in association with *P. lobata*, *P. lutea*, *Oxypora lacera*, *Pectinia lactuca*, *Horastrea indica*, *Blastomussa merleti*, *Gardinoseris planulata*, *Lobophyllia costata*, *P. stellata*, *G. pectinata*, *G. palauensis*, *L. purpurea*, *P. maldivensis*, *Cyphastrea microphthalma*, *H. microconos*, *Goniopora* sp., *E. gemmacea* and *D. heliopora*. The subordinate species include *F. pallida*, *F. speciosa*, *F. abdita*, *Lobophyllia hemprichii*, *P. damicornis* and *P. eydouxi*.

Owing to similarities in the taxonomic composition of the encrusting coral communities, the distinction between the relevant two subfacies in cores is not easy and has to be based on the texture and composition of matrices and the nature of secondary encrusters. The subfacies rich in terrigenous mud is considered to form at less than 10 m deep in protected environments, while the coralgal–foraminiferal subfacies is thought to deposit on deep fore-reef zones at depths greater than 20 m.

### 3.1.8. Skeletal rubble facies

Also known as coral rubble, coral gravel, skeletal rubble and coral rudstone, this facies is composed of a mixture of unsorted, angular to rounded clasts of corals, bivalves and coralline algae. It forms a prominent feature of most cored reef sections, irrespective of ambient water-energy conditions and morphological zones. The facies may occupy up to 60% of the total core volume in situations ranging from exposed reef margins to innermost backreef zones (Davies and Hopley, 1983; Johnson et al., 1984; Montaggioni, 1988; Tudhope, 1989; Kennedy and Woodroffe, 2000; Grossman and Fletcher, 2004). Since disintegration of coral framework is greatly influenced by the original shapes of coral colonies, branching acroporids, poritids and pocilloporids, and

foliaceous corals (*Montipora*, *Pavona* mainly) are the major contributors to rubble deposition. Skeletal pieces range between 0.5 and up to 3 cm in size; they can be either totally non-encrusted or heavily encrusted by coralline algae, associated with arborescent foraminifers and bryozoans, thus forming incipient rhodolites locally. These clasts may be severely bored by macrofauna (sponges, mytilid bivalves) and a variety of micro-organisms. Lithified, thick, laminar to columnar micrite crusts may locally encapsulate skeletal gravels, thus forming a solid framework and reducing initial porosity (Montaggioni and Camoin, 1993; Engels et al., 2004).

The gravel material generally is unconsolidated and may be supported by sandy and/or muddy matrix. In cores extracted from high-energy settings (i.e. exposed reef fronts and flats), when present, matrix is chiefly fine- to coarse-sand sized foraminiferal tests, micro-molluscs, debris of branched coralines, *Halimeda* plates, alcyonarian spicules, echinoid detritus and, possibly, brachiopods. The sediment bulk is of rudstone texture. In cores retrieved from low-energy areas (i.e. inner reef flats to backreef zones), the interclast matrix is composed of fine sand and locally mud. The mud content ranges from 1% to 80% of the total sediment (Johnson and Risk, 1987; Adjas et al., 1990; Smith et al., 1998; Kennedy and Woodroffe, 2000). The resulting facies varies from muddy sandy gravels to gravelly muds (unconsolidated floatstones). The mud fraction is either purely carbonate or mixed carbonate–terrigenous. Purely carbonate mud derives from disintegration of the skeletons of reef-dwelling organisms; it is mainly found on lagoonal platform reefs in mid-shelf situations (Smith et al., 1998) and within atoll lagoons (Adjas et al., 1990; Yamano et al., 2002). Mixed siliciclastic–carbonate mud sedimentation is restricted to nearshore reefs in inner shelf situations (Johnson and Risk, 1987; Hopley, 1982; Tudhope and Scoffin, 1994; Woolfe and Larcombe, 1998) and in mid-oceanic volcanic islands (Montaggioni and Faure, 1980; Cabioch et al., 1999a; Zinke et al., 2001; Engels et al., 2004).

The skeletal rubble facies is the equivalent of intertidal to subtidal storm-generated gravel sheets deposited at the surface of reef flats and prograding into backreef zones (Montaggioni and Faure, 1980; Hopley, 1982; Davies and Hopley, 1983).

### 3.1.9. Carbonate sand facies

Also known as skeletal, detrital, biogenic or bioclastic sand, this facies consists of gravelly (rudstones, grainstones) to muddy (packstones) sands derived from the disintegration of reef-dwelling organisms. Few works have been devoted to the biological composition of sand deposits collected by reef drilling; less than 10% of the Indo-Pacific core dataset available are documented from the viewpoint of total bioclastic components. This may be due to the low recovery rates associated with sand beds (1–30%; average: about 10%). Present in all cores, this facies is volumetrically one of the most significant to such a point that Hubbard et al. (1998) questioned the actual role of in-place framework in Holocene reef building. Indeed, even in cores retrieved from ocean-facing reef crests, sand intervals up to 5 m were penetrated (Webster et al., 1998; Cabioch et al., 1999a, 2003a,b; Grossman and Fletcher, 2004; Woodroffe et al., 2004).

The sand facies is widespread beneath a variety of reef zones. Drillholes dominated by sand accumulations have clearly sampled inner reef flat and backreef sites in which sands may occupy more than 80% of the total core volume (Figs. 3 and 4) (Marshall and Davies, 1982; Davies and Hopley, 1983; Marshall and Jacobsen, 1985; Montaggioni, 1988; Gray et al., 1992; McLean and Woodroffe, 1994; Cabioch et al., 1995, 1999a; Zinke et al., 2001). Conversely, sand beds range between 10% and about 50% in total volume beneath reef crests and outer reef flats (Fig. 2) (Easton and Olson, 1976; Marshall and Davies, 1982; Cabioch, 1988; Hopley, 1994; Montaggioni and Faure, 1997; Montaggioni et al., 1997a,b).

Sands can be unconsolidated to firmly lithified. Loose material mainly occurs in cores from leeward, protected areas. By contrast, densely cemented skeletal sands can be found in cores taken from windward reef margins and occasionally, from large coral patches irrespective of their location. These common features in all reef sites may be explained by the fact that the most favourable conditions for intergranular cementation seem to be strong water agitation, stabilization of sediments and occurrence of enclosed, intraframework micro-environments supersaturated with respect to calcium carbonate (Macintyre and Marshall, 1988).

The dominant components are scleractinians, coral-line algae, molluscs, benthic foraminifers and green algae (*Halimeda*). The subordinate producers are alcyonarians, ostracods, echinoderms, bryozoans, crustaceans and planktic foraminifers. Mineralogically, the relevant sands are largely composed of aragonite (50 to 80% of the total sand fraction) and magnesian calcite. The following biofacies can be distinguished in sandy sections on the basis of the major skeletal components and textural characteristics.

- coral–coralline algal (coralgal) rudstones/grainstones/packstones; they are ubiquitous, but occur chiefly in higher abundance (up to 50% of the total sand fraction) in the cores penetrating windward to leeward reef crests and outer reef flats. Associated bioclastic types derive mainly from benthic foraminifers and molluscs. Foraminiferal taxa are dominated by encrusting (Acervulinids, homotremids, victoriellinids) and hyaline forms (Amphisteginids, calcarinids and/or baculogypsinids) (Montaggioni, 1977; Webster et al., 1998; Kayanne et al., 2002; Cabioch, 2003; Yamano et al., 2003; Grossman and Fletcher, 2004).
- coral–molluscan grainstones/packstones; they are restricted mainly to cores from inner reef flats and adjacent shallower backreef zones. The subordinate biogenic particles come from corals, corallines, free-living benthic foraminifers (Amphisteginids, calcarinids, soritids, miliolids), alcyonarians and echinids (Johnson and Risk, 1987; Smith et al., 1998; Braithwaite et al., 2000; Kayanne et al., 2002; Yamano et al., 2002). Sand-sized debris of corals and molluscs may contribute to proximal fore-reef deposition in siliciclastic muddy sands (Cortés et al., 1994).
- molluscan–foraminiferal grainstones/packstones; they are dominating in the cored material from mid-to distal lagoonal areas, on barrier reefs and atolls. Amounts of molluscs (cerithid shells, fragments of ostreids and tridacnes) can exceed 35% of the total bioclasts; tests of free-living foraminifers (Nummulitids, miliolids, textulariids) represent 5–10% on average. The other carbonate producers are *Halimeda*, echinoderms, crustaceans, ostracods and occasional planktic foraminifers (Perrin, 1989; Cabioch et al., 1999a; Zinke et al., 2003a).

– *Halimeda* grainstones/packstones; they occur in cores from a variety of reef zones. The algal segments can concentrate in some sections, representing up to 35% of the total components beneath inner reef flats (Marshall and Davies, 1982; Engels et al., 2004) and backreef zones (Kayanne et al., 2002; Degauge-Michalski, 1993). However, *Halimeda* particles have been mixed usually with other skeletal material during reef accretion. Sequences from semi-exposed to sheltered fringing reefs may display a coralline algal–*Halimeda* packstone facies (Cabioch, 1988). In core sections from West to Central Pacific atolls, three successive *Halimeda*-dominated sand facies were identified in lagoonal settings, from the proximal, shallower to the distal, deeper parts: coral–*Halimeda* grainstones, *Halimeda*–foraminiferal (Nummulitids, miliolids) and *Halimeda*–molluscan packstones, respectively (Yamano et al., 2002).

Low-extent skeletal sand types were reported in sections from various reef environments: alcyonarian (spiculite) grainstones beneath reef flats and shallower backreef sites (Montaggioni, 1980; Konishi, 1981; Johnson and Risk, 1987; Braithwaite et al., 2000); coralline algal–molluscan packstones from shallower lagoon areas (Kennedy and Woodroffe, 2000); molluscan–ostracod packstone/wackestone in deeper lagoons (Zinke et al., 2003a,b).

The distributional patterns of the carbonate sand facies are distinctive, despite the significant overlaps in the skeletal associations. The many facets of pattern result undoubtedly from biogeographic and hydrodynamic mechanisms as well. The present analysis of sand facies zonation in cores allows two spatial scales of pattern to be coming out: a within-reef-zone scale and a between-region scale. Within a given reef zone, sand accumulations may be compositionally and texturally similar throughout the Holocene time interval. Particularly, in higher water-energy settings (i.e. windward reef margins), sand beds commonly are strictly composed of coralline grainstones (Montaggioni, 1977; Camoin et al., 1997; Webster et al., 1998; Cabioch, 2003; Yamano et al., 2003, submitted for publication). However, in most cases, there is broad constituent and grain-size variation over core length. For example, in a moderate water-energy, outer

fringing reef flat (Seychelles), Braithwaite et al. (2000) described a 26-m-thick sand-dominated sequence, changing upward from a medium- to fine-sized, coral–molluscan–alcyonarian grainstone, with nummulitids into a coarse sized coral grainstone/rudstone, rich in homotremids, victoriellinids and amphisteginids. Similarly, through a windward, mid-shelf reef flat (Australian Great Barrier Reef), Marshall and Davies (1982) drilled a 11-m-thick, upward fining succession, typified at base by *Halimeda* grainstones and at top by coralline packstones. The backreef environments also exhibit vertical changes in sand texture and composition. For instance, the core set extracted from Mayotte lagoon showed that Holocene deposits consist of upward coarsening, molluscan to molluscan–foraminiferal (nummulitids, amphisteginids) sands in shallower areas and, of upward fining, molluscan to molluscan–ostracod sandy muds rich in nummulitids and miliolids in deeper settings (Zinke et al., 2003a). Differences in Holocene bioclastic associations exist between the Indo-Pacific regions. These differences chiefly concern the distribution of *Halimeda* algae and benthic foraminiferal forms. *Halimeda* detritus seems not to have formed a distinctive sand facies during Holocene reef development in the islands from Western Indian Ocean (Montaggioni, 1977; Camoin et al., 1997; Braithwaite et al., 2000; Zinke et al., 2003a) and the Tuamotu archipelago (Pirazzoli and Montaggioni, 1986; Perrin, 1989). By contrast, the *Halimeda* facies is a common feature of the sediments accumulated in semi-exposed to sheltered reef flats and backreef areas from the Western Pacific (Marshall and Davies, 1982; Degauge-Michalski, 1993; Hopley, 1994; Chevillon, 1996; Kayanne et al., 2002; Yamano et al., 2002) and Hawaiian islands (Engels et al., 2004). Regarding the foraminiferal assemblages present in the cored sand beds, there are marked variations from ocean to ocean and between the Pacific regions. In the Indian Ocean, the reef flat sequences are characterized by the dominance of a *Homotrema*–*Amphistegina*–*Calcarina* subfacies, and the backreef sections by a Miliolid (*Triloculina*–*Quinqueloculina*)–*Textularia* subfacies (Montaggioni, 1978; Colonna, 1994; Braithwaite et al., 2000). Cores collected from reef flat zones in the Western Pacific exhibit a *Calcarina*–*Baculogypsina*–*Soritid* subfacies (Baccaert, 1987; Cabioch, 1988; Yamano et al.,

2001a, 2002; Kayanne et al., 2002). The latter seems to be missing in the central Pacific, where it is replaced by an *Acervulina*–*Amphistegina* association (Perrin, 1989). On the contrary, the backreef sections studied would present similar foraminiferal subfacies throughout the Pacific, namely a nummulitid (*Heterostegina* and/or *Operculina*)–miliolid (*Quinqueloculina*, *Triloculina*, *Spiroloculina*) association (Yamano et al., 2002; Perrin, 1989).

All of the skeletal sand facies described are the fossil counterparts of present-day reefal deposits. The patterns in the distribution of the biogenic component associations are relatively well documented from the Indo-Pacific (Lewis, 1967; Masse, 1970; Montaggioni, 1978; Flood and Scoffin, 1978; Gabri  and Montaggioni, 1982a,b; Montaggioni et al., 1986; Delesalle, 1985; Adjas, 1988; Cabioch, 1988; Chevillon, 1992; Piller, 1994; Zinke et al., 2001).

Generally, in high-energy environments, the composition of the sand fractions deposited only partially reflects the structure of the nearby carbonate-producing communities, since large-scale transport of sediments operate across the reef system. By contrast, in low-energy settings, the boundaries of the different sand facies coincide broadly with those of the communities (Montaggioni et al., 1986; Piller, 1994). Similarly, numerous studies were focused in particular on reef biozonation based on the presence or absence of dominating larger foraminifers within surface sand material (Todd, 1960; Montaggioni, 1981; Debenay, 1985; Hallock and Glenn, 1986; Montaggioni et al., 1986; Venec-Peyr , 1991; Bicchi et al., 2002; Yamano et al., 2002; Langer and Lipps, 2003). These indicate that the thanatocoenoses from reef margins, flats and shallower backreef sites are a mixture of allochthonous and in-place tests, while the deeper parts of lagoons mainly contain autochthonous microfaunal associations. As a consequence, these facies can be used as paleoenvironmental indicators. The corallgal facies, plus the foraminiferal subfacies dominated by encrusting forms, amphisteginids and/or calcarinids, are typical of windward reef margins and outer reef flats; they are diagnostic of the shallowest and highest energy settings. The coral–molluscan sand type, in association with the amphisteginid–calcarinid or the baculogypsinid–calcarinid–soritid variants is transitional; it is found mainly along inner reef flats and adjacent lagoonal areas at depths

rarely exceeding 10 m. Contrary to the corallgal, in the coralline algal–molluscan facies, calcifying algae are represented mainly by branched and articulated forms: this facies occupies the same reef zone as the coral–molluscan one. The molluscan–foraminiferal facies, rich in *Heterostegina* and/or *Operculina*, *Quinqueloculina*, *Triloculina* and *Spiroloculina* is extending over the deeper backreef bottoms; it can be overlapped locally by the *Halimeda* facies. The molluscan–ostracod assemblage appears to signify restricted, deeper lagoonal settings. The four mollusc-rich sand types are interpreted as lower energy facies. The alcyonarian spiculite facies, that results from the decay of octocorallian tissue after death, occurs in situ deposition, commonly on shallow subtidal, soft sedimentary bottoms (less than 10 m deep) (Montaggioni, 1980; Konishi, 1981); it is indicative of moderate agitation in shallow-water areas.

#### 3.1.10. Carbonate mud facies

Also referred to as lime mud, carbonate mudstone, this facies consists of fine-grained, carbonate-dominated sediment (grain size < 63 µm). It is encountered in cores from a variety of reef zones as either unconsolidated silt–mud deposits or lithified, microparticulate to micritic limestones (in the sense of Bathurst, 1971).

Cores from sheltered, inner reef flats and backreef zones generally contain loose fine-grained matrix that frequently encases isolated coral colonies or gravels (floatstones) and/or skeletal sands (wackestones). The relationships between water agitation and amounts of fine carbonates in cores remain relatively constant throughout the sediment piles. Beneath reef flats, the sediment is composed of less than 10% carbonate mud (Johnson and Risk, 1987; Yamano et al., 2001a; Braithwaite et al., 2000; Kennedy and Woodroffe, 2000). Higher contents of mud (50% to 98% of total sediment) are found generally in cores from the deeper parts of barriers and atolls (Smith et al., 1998; Zinke et al., 2001). During infilling of inner reef zones, there have been abrupt changes in the volume of fine-grained material deposited that tends to regularly increase up the cores (transition from floatstone–wackestone to mudstone textures) (Johnson and Risk, 1987; Zinke et al., 2001), probably as a result of the rapid vertical accretion of outer reef crests that thus have protected the adjacent inner reef zones and



prevented the potential skeletal material to be exported.

The origins of unconsolidated carbonate oozes in Holocene reef sequences are poorly documented. This relates presumably to the difficulty of both recovering fine-grained fractions even by vibrocore and recognizing the initial nature of a material highly susceptible to syndepositional alteration. The respective influence of the two basic mechanisms invoked for explaining the origin of the lime mud in reefs (Bathurst, 1971) is not known in the case of the Indo-Pacific sequences: has the mud been mostly inorganically precipitated or mostly derived from the breakdown of skeletal material? The only way of addressing this question is to investigate the composition of surficial, present-day fine-grained carbonates.

It appears that in the Indo-Pacific reefs, silt and mud are dominantly skeletal in origin (Scoffin and Tudhope, 1985; Tudhope and Scoffin, 1986; Ellis and Milliman, 1985; Adjias et al., 1990; Zinke et al., 2001). Low amounts of mud may be chemogenic, deposited as aragonite or magnesian calcite crystallites in restricted lagoonal environments seasonally affected by carbonate supersaturation (Adjias et al., 1990) or in arid, subtidal to intertidal, backreef flats (Purser, 1973).

Lithified lime mud is found mainly in cores from high-energy settings (i.e. windward reef margins, outer reef flats) where early marine cementation is widely operating in cavity-infilling deposits within reef framework; the densely lithified framework sections form “pavement limestones” (Macintyre and Marshall, 1988). Apart from the clearly identified silt-sized bioclasts, the origin of the other components present in the carbonate microsparites and micrites still remains controversial (Macintyre and Marshall, 1988). In a number of Postglacial sections from the Indo-Pacific, the laminar to columnar micritic crusts observed within reef framework were interpreted as microbially induced (stromatolite-like structures) (Camoin et al., 1999) rather than a physico-chemically deposited product.

### 3.2. Reef anatomy

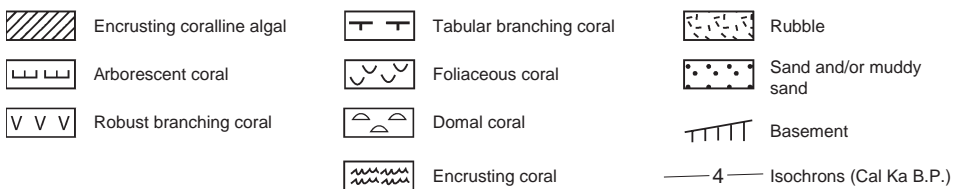
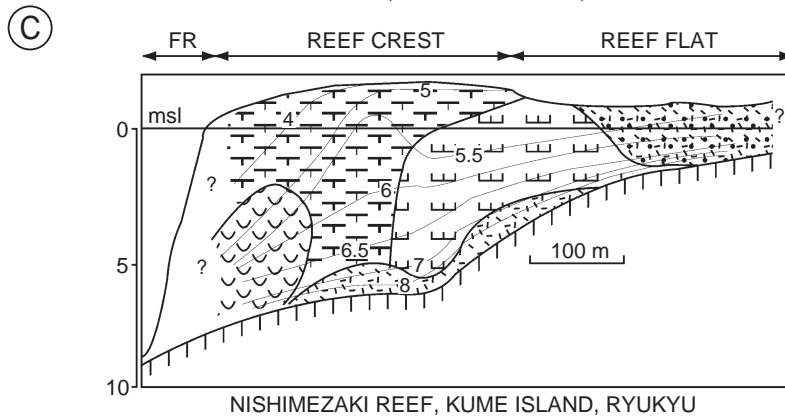
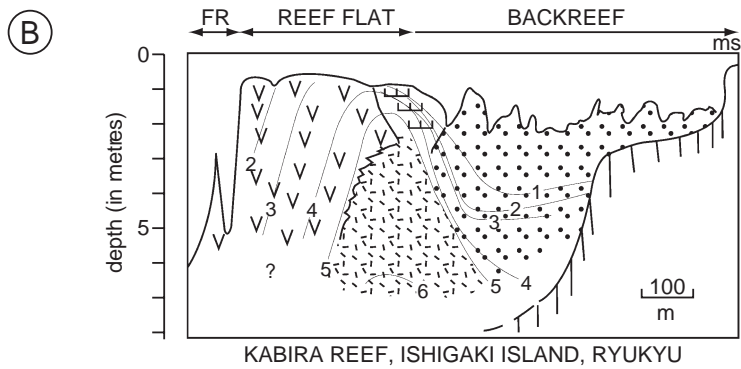
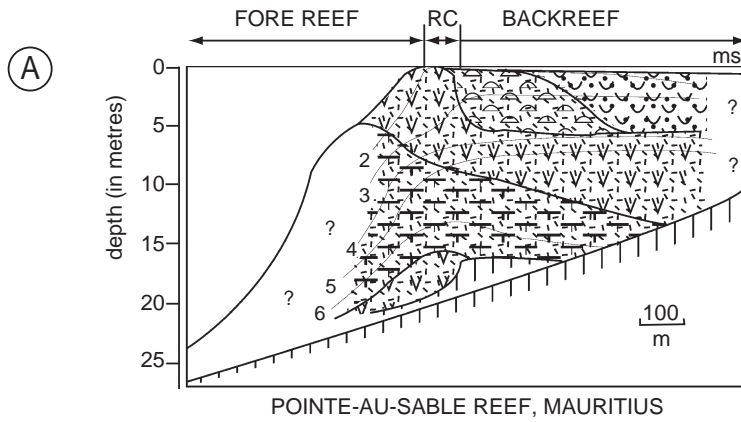
#### 3.2.1. Facies association

Although reefs are considered generally to be complex mosaics of facies (James and Bourque,

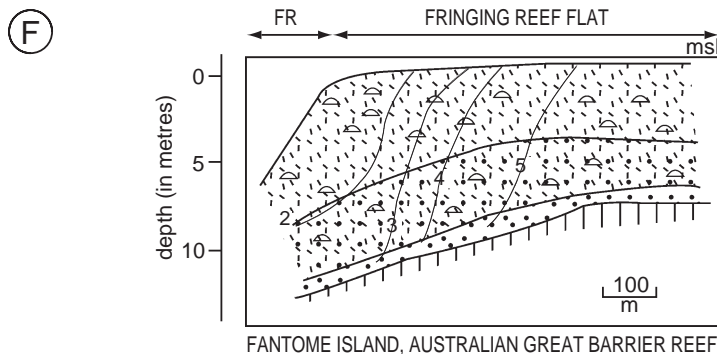
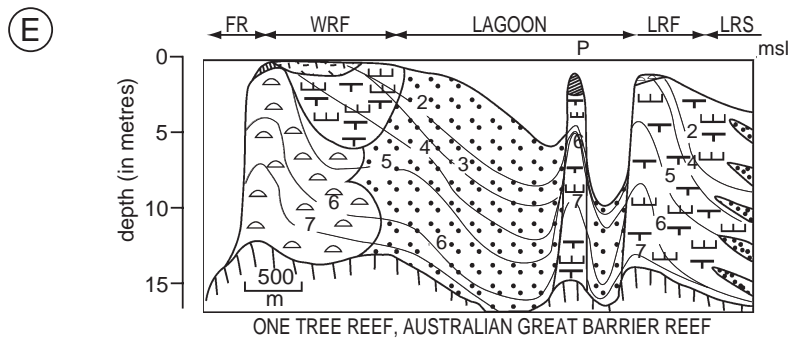
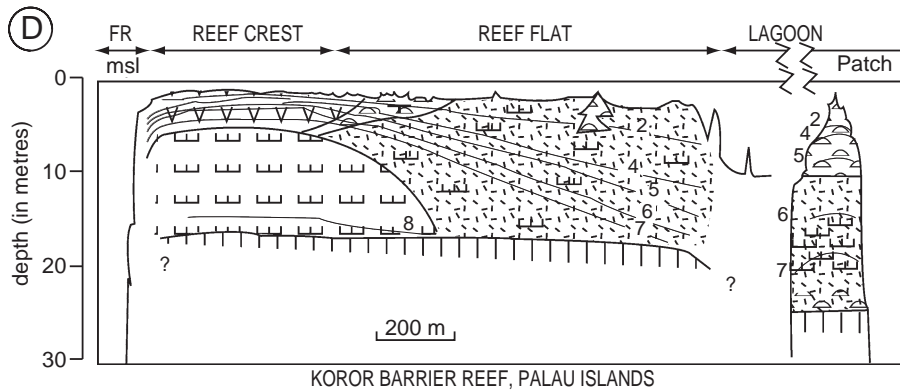
1992), models of reef development are possible to be defined on the basis of the distributional patterns of the facies both vertically and laterally through the reef system pile. The reconstructions of reef anatomy herein are largely based on core-transects.

Drilling showed that the internal structure of reef margins is typified by three types of facies association. The first type relates to sequences that comprise framework facies of uniform composition from base to top. These sequences are encountered generally beneath exposed to sheltered reef crest/flat zones that started to accumulate not prior to 10–9 ka BP. Single framework facies dominate in protected inshore areas. This model is illustrated by fringing reefs in South Thailand (Tudhope and Scoffin, 1994), Fantome island (Johnson and Risk, 1987) (Fig. 6F) and Punta Islotes (Cortés et al., 1994) (Fig. 6L). In turbulent offshore waters, facies of homogeneous composition are found locally. For instance, the high-energy reef flat sequence at Toliara (Fig. 2) is totally composed of a single robust-branching *Acropora* facies, associated with beds of unconsolidated sand and rubble mainly derived from acroporid branches (Camoin et al., 2004). Similarly, on Lord Howe island, beneath the relatively protected reef crest, the sediment pile consists of an arborescent *Acropora* framework locally mixed with detritus dominated by pieces of acroporid branches (Kennedy and Woodroffe, 2000) (Fig. 6G). Fringing reef flats at Réunion, Ishigaki, Poum, Orpheus, Molokai and Cocos are additional examples of this first type of facies succession.

The second type relates to the stacking-up of two distinct framework facies in a given core. In most cases, a bed of deeper water, lower energy coral community is overlain by a shallower, higher energy coral (or coralgall) assemblage. From the margins subjected to exposed or semi-exposed conditions, e.g. Mayotte, Mahé as Indian Ocean examples, and Palau, Guam, Mangaia, Yonge, windward margin of One Tree, Mamié, Moorea barrier, Mururoa as Pacific examples (Figs. 2, 3 and 6D,E), the base of the sequences consists of domal poritid/faviid, arborescent or tabular acroporid frameworks that represent former buildups started at depths of around 10 m if considering the present thickness of the sequences. The overtopping unit is composed of robust acroporid/pocilloporid assemblages encrusted by coralline

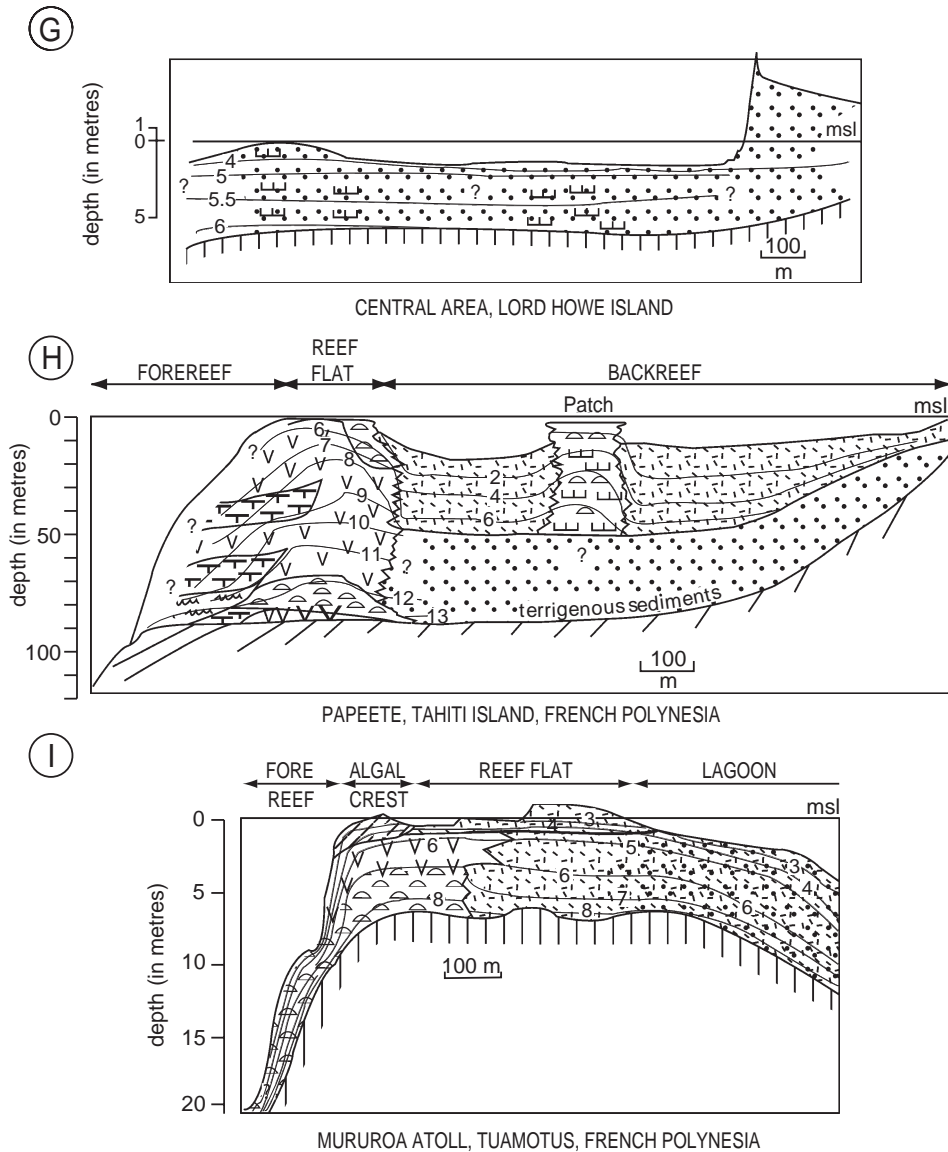






algae (*Hydrolithon* mainly) or by algal pavements, both representing the shallowest and highest energy facies. Within the cores from more protected reef rims (Abrolhos, Kume, inner margin of One Tree, fringing reefs at Moorea and Oahu) (Figs. 3 and 6E), the lower sections are composed of arborescent acroporid or poritids, or of foliaceous montiporid/agariciid/dendrophylliid assemblages that relate to the lowest energy and/or deepest facies, initially settled at depths not exceeding 15 m. The upper sections are dominated by domal poritid/faviid or tabular acroporid facies.

The third type of facies association is observed mainly from the reef crest sites that recorded the longest development history, i.e. the last 14 to 23 ka. At Kwambu, Tasmaloun and Tahiti (Fig. 2), the stratigraphic pattern is characterized by recurrent alternations of shallower, higher-energy and deeper, lower-energy frameworks. For example, the reef crest pile from the Tahitian barrier is made up of 1–10 m thick, alternating beds of shallower *A. gr. robusta* and deeper tabular *A. gr. cytherea*, arborescent *A. clathrata*, domal *Porites* spp, encrusting *P. lichen* and



arborescent *P. nigrescens* (Cabioch et al., 1999a) (Fig. 6H). This type of facies transition is also exemplified by fringing reefs, at Mauritius (Fig. 6A) and Kume (Fig. 6C).

Facies transitions are observed also within the backreef sediment piles. Scoffin and Tudhope (1988) established a predictive model of complete lagoonal sedimentation from their study on a small-size, enclosed lagoonal platform reef system (Davies Reef, mid-shelf Great Barrier Reef). They showed that the

relevant shallowing-upward sequences start with a subtidal unit consisting of in situ-produced coral rudstones, overlain by skeletal floatstones/wackestones in which sorting, mean grain size and amount of reef flat constituents increase near the sea surface; the upper part of the sequences is an intertidal to supratidal unit presenting a clear differentiation according to exposure to water energy: coral shingle rudstones to windward, grainstones to leeward. Similar subtidal sequences composed of fining-

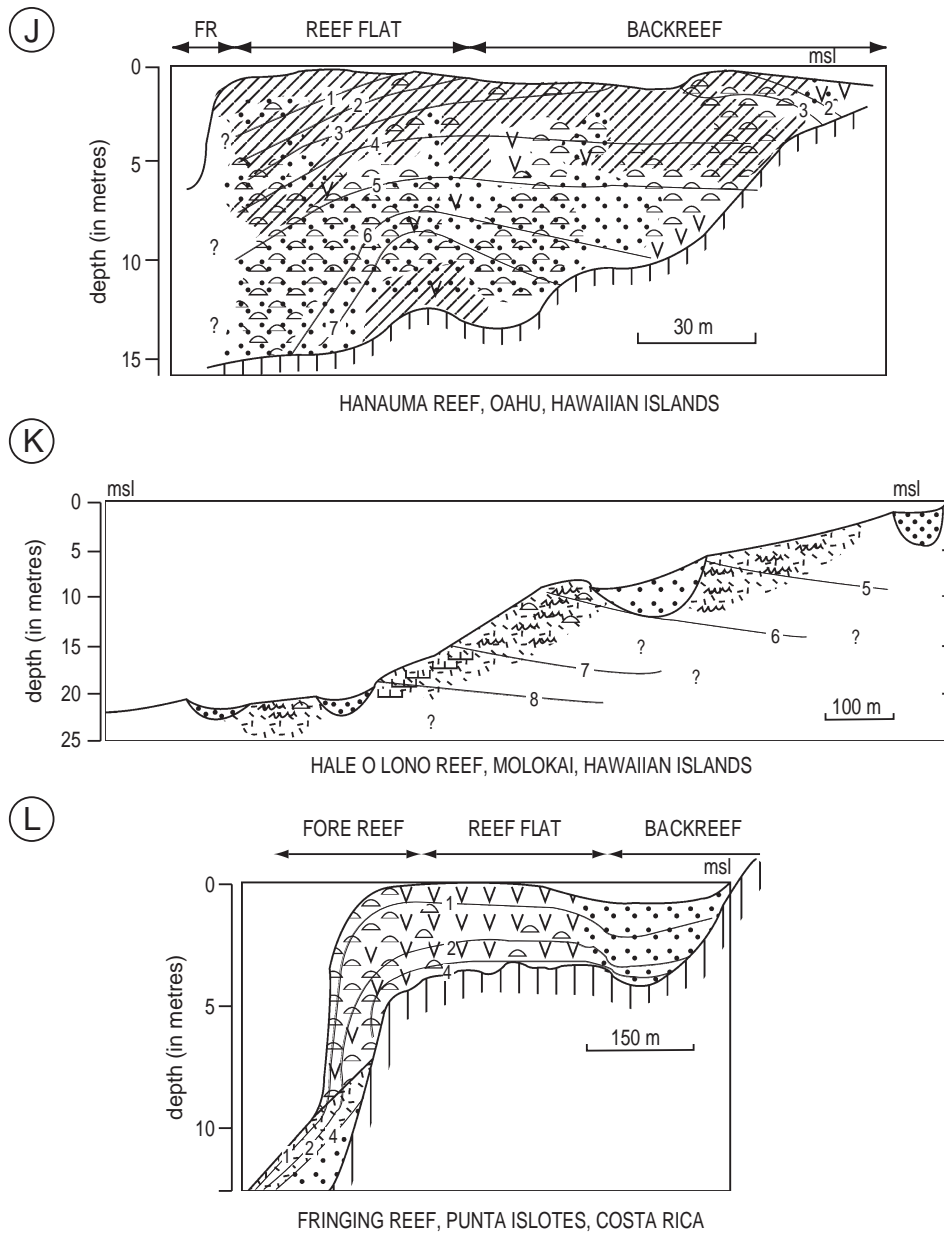


Fig. 6. Facies and anatomy patterns of selected Indo-Pacific reef systems: (A) fringing reef, La Pointe-au-Sable, Mauritius island, Indian ocean (Montaggioni and Faure, 1997); (B) fringing reef, Kabira Ishigaki island, Ryukyus, Western Pacific (Yamano et al., 2001b); (C) fringing reef, Nishimezaki, Kume island, Ryukyus (Takahashi et al., 1988); (D) barrier reef system, west off Koror islet, Palau island, Western Pacific (Kayanne et al., 2002); (E) mid-shelf platform reef, One Tree, Central Great Barrier Reef (Marshall and Davies, 1982); (F) inner-shelf fringing reef, Fantome islet, Central Great Barrier Reef (Johnson and Risk, 1987); (G) fringing reef, Central area, Lord Howe island, Tasman Sea, Western Pacific (Kennedy and Woodroffe, 2000); (H) barrier reef system, Tahiti island, Central Pacific (modified from Cabioch et al., 1999a); (I) outer reef margin, Mururoa atoll, Tuamotu archipelago, Central Pacific (Ehren, 1996); (J) fringing reef, Hanauma Bay, Oahu island, Hawaii archipelago, Central Pacific (Easton and Olson, 1976); (K) fringing reef, Hale O Lono, Molokai island, Hawaii archipelago, Central Pacific (Engels et al., 2004); (L) fringing reef, Punta islotes, Costa Rica, Eastern Pacific (Cortés et al., 1994).

upward sediments separated by in-situ low-energy corals and possibly grading into coarser-grained material closer to sea level, were described from different mid-shelf reef platforms on the Great Barrier Reef (Marshall and Davies, 1982; Davies and Hopley, 1983; Smith et al., 1998; Webster, 1999). Although the facial and textural characteristics of backreef sediment piles still remain poorly documented, the core data available seem to indicate that this pattern of lagoonal deposition is a common feature in different reef system types, at least regarding the subtidal beds, and irrespective of the size and area of the back reef environments (Fig. 4): fringing reefs (for instance Mauritius, Ishigaki, Rattelsnake, Lord Howe), oceanic atolls and platforms (Chesterfield, Mataiva). An alternative model of lagoonal sedimentation was reported by Zinke et al. (2003a) from the study of the large lagoon of Mayotte island. The facies successions show significant lateral variability (Fig. 7). Three depositional sequences were identified on the basis of different environmental settings. In the proximal lagoonal setting (10–35 m deep), the sequence includes a basal terrigenous, mollusc-rich mud overlain by mixed terrigenous/carbonate foramol wackestones, then carbonate coral–foramol packstones/rudstones. In the deeper, mid-lagoonal areas (40–65 m deep), the succession is as follows from base upward: terrigenous mud/wackestones rich in molluscs and ostracods, mixed terrigenous/carbonate molluscan–echinoidal wackestones, carbonate foramol packstones. From the distal lagoonal areas (10–45 m deep), the sequence consists of foramol mudstones/packstones capped by coral packstones/rudstones.

### 3.2.2. Reef anatomy versus structure of coral communities

From the foregoing, it is clear that the reef anatomy of Indo-Pacific reef systems reflects complex histories of accretion during the rise of sea level. Reef development expresses two distinct episodes in terms of major depositional processes: a transgressive episode from 19 to about 7 ka BP, and a stillstand episode for the past 7–3 ka (Davies and Montaggioni, 1985; Montaggioni, 2000). The transgressive successions are the products of a series of dominantly aggrading, shallowing-up sequences. The still-stand interval was characterized rather by lateral reef-margin accretion and backreef infilling.

The rise of sea level resulted in a progressive increase in water depth and change in wave power on the inundated substrates. The local capacity of reef growth to compensate for the increase in accommodation and to resist to water-energy space was determined by the structure of coral communities. Faster-growing, branching assemblages were generally able to keep pace or to prograde, while slower-growing, domal and laminar populations commonly retrograde or drowned. Thus, the vertical distribution of primary framework within reef rim piles reflects the response of communities to changes in rates of sea-level rise and hydrodynamic energy through changes in depositional processes and/or growth styles. Homogeneous composition of framework facies within a given sequence may indicate the persistence of ambient conditions from the earlier stages of coral colonization to coral growth at the stillstand. In particular, in sheltered inner-shelf settings where inimical conditions are operating (e.g. high terrigenous input, abnormal salinities), only the most tolerant coral assemblages, such as those dominated by poritids, faviids or pocilloporids, can survive. By contrast, the repeated abrupt facies replacements are interpreted as reflecting lateral displacements of coral communities across the drilling sites. The overlapping of robust acroporid-dominated framework by a deeper coral assemblage relates probably to a retrograding event, while the seaward displacement of the robust acroporids over deeper water corals express progradation. This suggests that rapid changes in environmental parameters have disturbed reef aggradation during the deglacial transgression. In particular, any decrease in the rate of sea-level rise has resulted in progradation of shallowest corals and, conversely any acceleration has favoured retrogradation of deeper corals. Such low-amplitude, high-frequency sea-level pulses may have controlled reef aggradation through successive keep-up/catch-up cycles. Finally, in so-called keep-up reefs, the cycles have been encapsulated as an overall keep-up signature.

In sequences from semi-exposed to sheltered mid-shelf and inner-shelf settings, domal, arborescent and tabular-branching coral assemblages dominate over other coral communities and may locally form uniform facies. They generally indicate aggradation controlled by a catch-up growth mode.

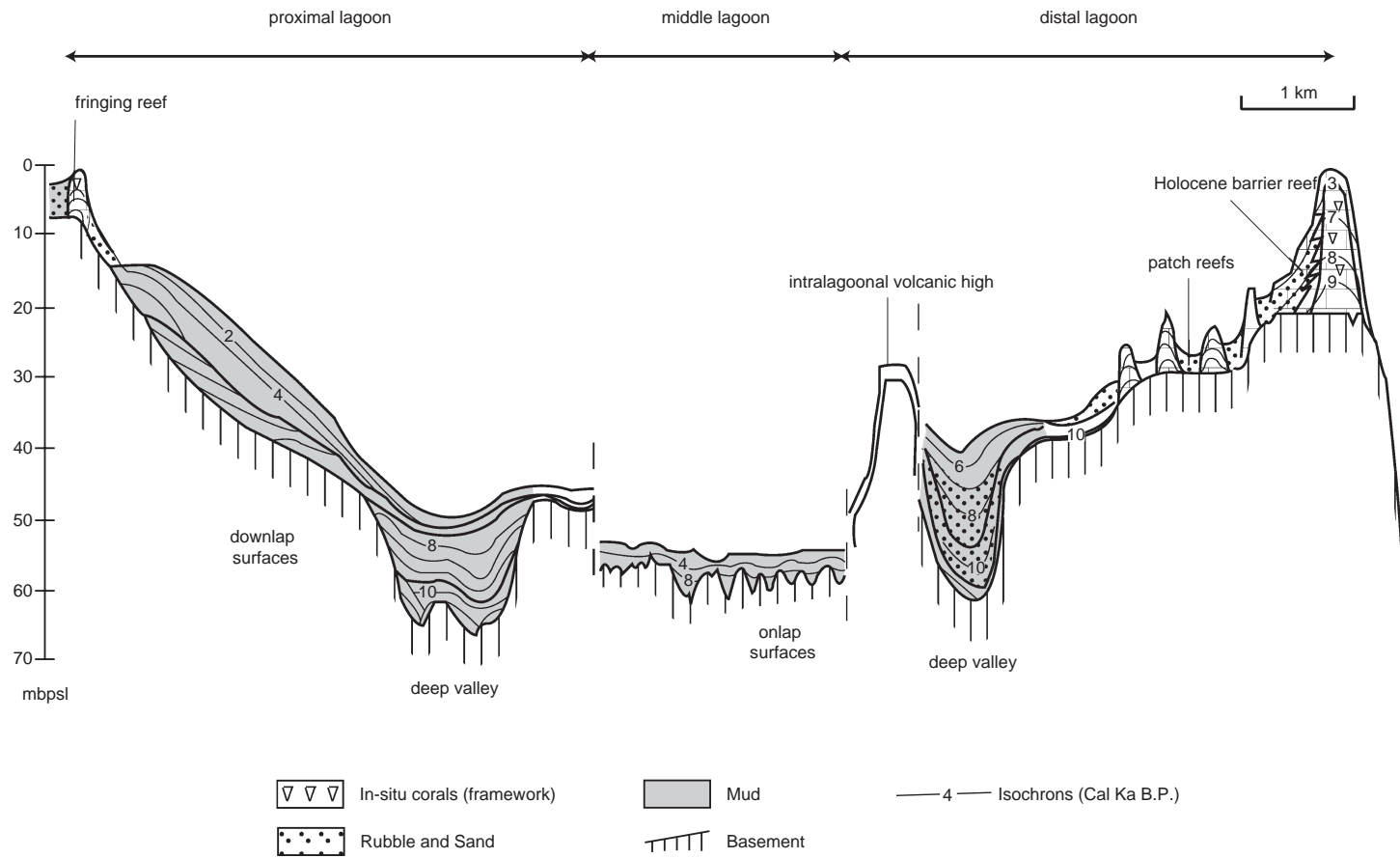


Fig. 7. Generalized seismic-based cross-section of the barrier reef system at Mayotte island, Western Indian Ocean. Basal unconformity, depositional systems and sequences boundaries of Postglacial age are indicated (from Zinke et al., 2001).



There is evidence of a link between the structures of coral communities and the degree of reef development. The attributes of the community, species composition and abundance, colony size frequency distribution, individual life expectancy and succession, all influence reef accretion, either within the in-situ framework or in detrital deposits (Davies, 1983; Kleypas, 1996; Van Woesik and Done, 1997). In the southern Great Barrier Reef, non-reefal coral communities (in the sense of Buddemeier and Hopley, 1988) and incipient reefs that are submerged and lack a defined reef flat (Hopley et al., 1989), are dominated by encrusting and foliaceous growth forms. The main reef-builders (branching and large massive forms) and the major rubble producers (branching corals) are rare or absent. If the living community provides an adequate guide to coral palaeoassemblages and palaeoenvironmental conditions throughout the Holocene, it should not be surprising that reefs have not developed. High population turnover, low settlement densities and high rates of skeletal breakdown combined with high turbidity prevented substantial accretion. These structures, interpreted as “turned off” are largely composed of detritus. By contrast, fringing reefs with a well-defined reef flat are regarded as “turned on” (Kleypas, 1996). These consist predominantly of branching acroporids that have provided framework and rubble. If maintained throughout the Holocene such conditions, reflecting high recruitment densities and long colony life expectancies, would have promoted substantial net growth.

Furthermore, the results from drilling underline the dominance of growth framework in exposed, ocean-facing reef margins, contrasting with those in sheltered, inner-shelf tracts. This gross facies distribution, where the relative proportions of framework and detritus are related to hydrodynamic energy, seems to reflect a general rule. In areas subject permanently to wave energy, the margins of outer and mid-shelf reefs and ocean-facing fringing reef fronts consist of an in-situ, interlocking, primary growth frame, in the sense of Hubbard et al. (1998, 2001). The framework forms 40–50% of the total core volume (Montaggioni, 2001). In the highest water-energy situations, particularly within hurricane-swept regions, reef margins consist mainly of a secondary framework (in the sense of Hubbard et al., 1998), reflecting redeposition and algal encrustation of displaced corals. In these, the

total framework facies (primary and secondary) may comprise more than 60% of the cored material. In the lowest energy settings, that may or not be subject to hurricanes and tropical storms, reefs can best be described as detrital piles trapping only scattered corals (Davies and Hopley, 1983; Montaggioni, 1988; Kleypas and Hopley, 1993; Cabioch et al., 1995; Braithwaite et al., 2000; Yamano et al., 2001a). Coral colonies typically comprise less than 20% of the total core. Thus, in the Indo-Pacific, reef anatomy appears to be partly dependent on the interplay between fairweather hydrodynamic conditions and extreme storm events. It can be related to end-members of a hydrodynamic-controlled spectrum ranging from framework to detritus.

In addition to controlling the large-scale architecture of reef systems, hydrodynamic energy may provide small-scale constraints on reef geometry through framework composition. Operating in parallel with sea-level change, wave agitation may be responsible for limiting reef growth and, as a consequence, for controlling the thickness of reef sequences. In the high Hawaiian Islands, Grigg (1998) observed that Holocene reefs 10–15 m thick and dominated by domal *Porites* occur only in wave-sheltered locations. In wave-exposed settings, accretion is represented by less than 1 m of living corals overlying a variety of antecedent foundations.

### 3.2.3. Reef anatomy versus depositional patterns

Kennedy and Woodroffe (2002) defined six models of Holocene fringing reef development based on the use of isochrons to reconstruct the successive stages of accretion. Revisited on the basis of dominant depositional patterns, this classification can be widened to the different reef morphologies known in the Indo-Pacific domain (i.e. fringing and barrier reefs, platform reefs and atolls). All of them can be accommodated to the following main four anatomy types of reef systems (Fig. 8).

In the first type, deposition within lagoons operated at rates close to those of the adjacent reef rim allowing the different parts of the system to reach sea level at the same time (Fig. 8A). There was a fine balance between reef rim growth through aggradation and backreef accumulation through onlapping. The dominant process refers to balanced aggradation/onlapping (model A in the sense of Kennedy and

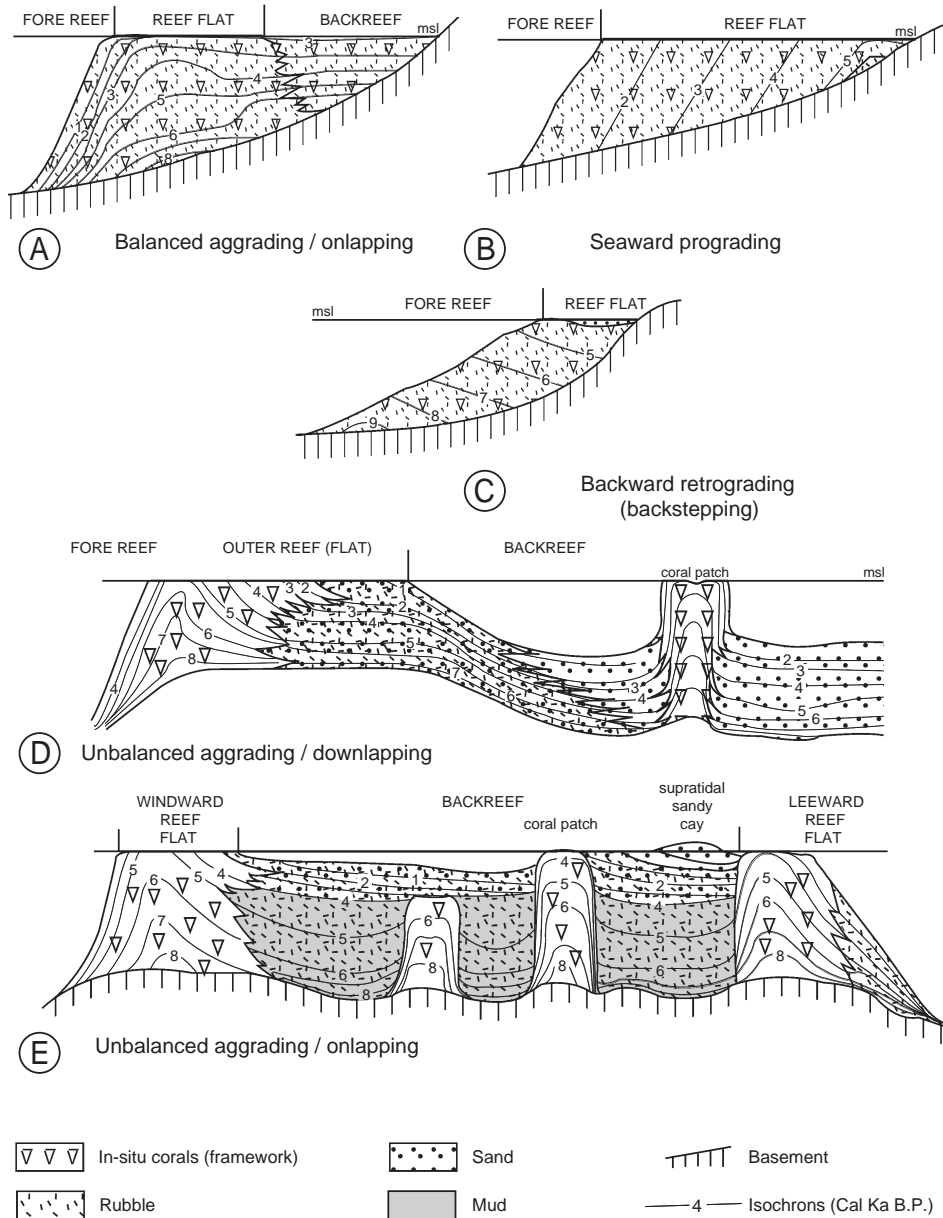


Fig. 8. Generalized models of reef system anatomy defined on the basis of dominant depositional processes. (A) balanced aggrading-onlapping model; (B) seaward prograding model; (C) back-stepping model; (D) unbalanced aggrading/downlapping model; (E) unbalanced aggrading-onlapping model (revisited from Kennedy and Woodroffe, 2002).

Woodroffe, 2002). Fringing reefs at Mauritius (Montaggioni and Faure, 1997), Kume (Takahashi et al., 1988), Lord Howe (Kennedy and Woodroffe, 2000), Hanauma and Kailua Bays (Easton and Olson, 1976; Grossman and Fletcher, 2004) are examples of this model (Fig. 6A,C,G,J).

The second type relates to reef systems that developed seaward by lateral accretion of the forereef. They started to grow in settings where vertical accommodation space was limited or missing. Such Holocene reefs mostly developed close to the shore when sea level was around its present-day position

(Fig. 8B). Referring to models B and C recognized by Kennedy and Woodroffe (2002), they are common in inner shelf sites. Fringing reefs at Phuket (Tudhope and Scoffin, 1994), Fantome (Johnson and Risk, 1987) (Fig. 6F) and Orpheus islands (Hopley et al., 1983), and at Hikauhi, Molokai island (Engels et al., 2004) are typical of seaward prograding systems. Lateral backward accretion can occur episodically; in this case, the reef develops through coalescing of offshore patches onto the forereef zone from framework growth and/or sediment infill (model D of Kennedy and Woodroffe, 2002). Such a pattern of progradation was reported from fringing reefs at Mahé, Seychelles (Lewis, 1968; Braithwaite et al., 2000) and from barrier reefs at Toliara, Madagascar (Weydert, 1973).

In the third type of reef anatomy (Fig. 8C), reef initiation occurred below modern sea level at depths where the rates of reef aggradation during the early Holocene times were not efficient enough to fill up available accommodation space. The entire reef body therefore caught up to sea level through successive accretion centres stepping backward; it finally comprises a series of superimposed retrograding units. The fringing reef at Hale O Lono, Molokai island (Engels et al., 2004) (Fig. 6K) is an example of this architecture. The landward migration of the accretion centres is assumed to result from the specific adaptation of coral communities to local environmental changes. Changes in the rate of the rising sea level would have contributed to the termination of former accretion centres by progressively modifying the local current regime and, thus, causing the reduction of wave sheltering at the reef site. Coral communities would be forced to displace landward to find habitat suitable for sustained growth (Engels et al., 2004).

The fourth type of reef anatomy relates to reefs that comprise a well-developed, outer rim enclosing a depressed backreef area forming through multiple depositional processes. In backreef settings, the differences observed between the stratigraphic columns express the existence of local depositional modes, irrespective of the nature of sediment sources (carbonate or terrigenous). As pointed out by Smith et al. (1998), these differences are dictated probably by the timing of reef rim growth and its relationships across the reef system to sea-level position during

the course of sea level. During the earlier times of inundation, the backreef zones probably were open systems in which sediment accumulation was continuously disrupted by water circulation. Except in the deepest depressions incising the inner floors, the current regime promoted coral settlement and winnowing of finer grains and, as a result, deposition of coral gravel. As the reef margins developed vertically, water agitation within the backreef areas decreased progressively and the sedimentation style changed. This resulted in the grading of sediments into sand and/or muddy sand. After the transgressive phase ended around 7 ka BP, most of reef rims that developed through keeping pace with sea level, formed protecting barriers, thus favouring trapping of sediments within backreef environments. The fact that each individual backreef zone responded with different growth styles and sedimentation rates to local environmental constraints, such as size, depth and topography of the antecedent buckets or basins, proximity to sediment source areas, suggests lagoonal deposits encapsulate local rather than global events. This explains why the main phase of backreef infilled started at around 7.6 ka BP (Marshall and Davies, 1982; Davies et al., 1985; Pirazzoli and Montaggioni, 1986; Smith et al., 1998; Cabioch et al., 1999a; Yamano et al., 2001a; Kennedy and Woodroffe, 2002) (Fig. 6E,H). Two distinct patterns of bed termination in backreef settings can be inferred on the base of isochron distribution or seismic profiles. Sediment sheets can have accumulated as downlapping clinofolds (Fig. 8D) at the inflection of slope between the rim-lagoon transitional slope and the lagoon floor settings. The isochrons delineate a series of talus cones deposited by by-passing of sediments from the windward margins or inshore terrigenous environments towards the centre and the distal parts of the lagoons or along the outer slopes of leeward reef margins (Figs. 6B,D,E,I, 7). Backreef deposition can also occur through onlapping beds caused by termination of gently dipping to subhorizontal sheets against steeply dipping antecedent substrate or flanks of intralagoonal coral patches; the isochrons are parallel to the lagoon floor surface (Figs. 6H,I, 7, 8E). Correlative to backreef infilling, vertical accretion of the reef margins occurred until the margin top reached present sea-level position; then, the former prograded backward over early

deposited backreef sediment beds (Figs. 6D,E,I; 8D,E). According to their initial depth and width and the sedimentation rates, the backreef bottoms either remained under subtidal conditions, filled up completely or passed through intertidal to supratidal deposits (Fig. 8D,E). As a whole, the backreef zones of fringing and narrow barrier reefs (less than 2 km wide), small-sized reef platforms and atolls (less than 50 km<sup>2</sup> in area) that initiated from antecedent substrate at depths of about 20–25 m below present sea level are in process of complete infilling through backward progradation and onlapping. This status is observed on fringing reefs at Mahé (Braithwaite et al., 2000), Ningaloo (Collins et al., 2003), Kabira (Yamano et al., 2001a; Fig. 6C), Torres Strait (Woodroffe et al., 2000), on mid-shelf platforms (Marshall and Davies, 1982; Scoffin and Tudhope, 1988; Smith et al., 1998). It may be inferred in the case of the barrier reef at Moorea (Montaggioni, 1988) and some atolls, e.g. Kayangel (Yamano et al., 2002), Mataiva (Pirazzoli and Montaggioni, 1986), Rakahanga (Gray et al., 1992). On reef platforms, barriers and atolls overlying deeper antecedent substrate, the lagoons commonly, particularly those nearly enclosed to enclosed, are partially filled by sediments and display a typical basin topography as emphasized by Zinke et al. (2003a,b). The main operating depositional process has been downlapping. This occurs in the majority of barriers and atolls, for instance, at Mayotte (Zinke et al., 2001), Palau (Kayanne et al., 2002), Middleton (Woodroffe et al., 2004), Enewetak (Tracey and Ladd, 1974), Tarawa (Marshall and Jacobsen, 1985), Tahiti (Cabioch et al., 1999a), Mururoa (Perrin, 1989) and in outer-shelf ribbon reefs (Davies and Hopley, 1983).

### 3.3. Rates and modes of reef accretion

Numerous studies reported measurements of reef accretion through the Holocene (Tables 3–7). Estimates of linear accumulation rates were made using radiometric dating of cores (Davies, 1983). These consist of measuring the thickness of deposited carbonates (i.e. in-situ coralgal assemblages, framework and detrital material) vertically and laterally, divided by the time interval over which these sediments have deposited as determined by dating. As pointed out by Davies (1983), calculating accretion

rates requires that reworking and displacement of sediments have not severely disturbed the original depositional fabric. In other respects, Blanchon and Blakeway (2003) considered that the rates of vertical accretion could be overestimated as a consequence of artefacts of reef coring. Accretion rates calculated from dating therefore should represent approximate values (Fig. 9).

In framework-dominated reefs, the total variation in vertical accretion rates ranges between 1 to up to 30 mm year<sup>-1</sup>, with a modal rate of 6–7 mm year<sup>-1</sup> (Table 3). High rates of aggradation have been recorded from both robust-branching and domal coral communities. For instance, the vertical accretion rates of high-energy, robust-coral facies may have reached 15 mm year<sup>-1</sup> in Mauritius (Montaggioni and Faure, 1997) and 14 mm year<sup>-1</sup> in Panama (Glynn and Macintyre, 1987). Similarly, lower energy, domal poritid assemblages grew upward at approaching rates of up to 13 mm year<sup>-1</sup> at Tahiti and Molokai island (Engels et al., 2004), and at rates greater than 16 mm year<sup>-1</sup> in Lord Howe Island (Kennedy and Woodroffe, 2000). However, it is noteworthy that the highest rates measured in the Indo-Pacific (up to 20 mm year<sup>-1</sup>) coincide with the development of higher porosity framework laid down by tabular and arborescent acroporid communities; rates of 20 to 30 mm year<sup>-1</sup> have been reported from gracile acroporid-rich sections beneath both the barrier reef flat drilled in Tahiti (Montaggioni et al., 1997a,b) and in Palau (Kayanne et al., 2002). Lower rates have been produced by foliaceous and, presumably, encrusting coral frameworks (maximum rate: 6 mm year<sup>-1</sup>) and by coralline algal facies (maximum rate: 3 mm year<sup>-1</sup>). Accretion rates may be highly variable within a single coral assemblage, as illustrated by the robust-branching facies on the semi-exposed west coast of Mauritius where accumulation rates range from 1 to 15 mm year<sup>-1</sup> (Montaggioni and Faure, 1997).

It has commonly been suggested that the nature of framework facies controls accretion rates, but this idea has proved controversial (for instance, see Davies and Hopley, 1983; Davies et al., 1985; Davies and Montaggioni, 1985; Cabioch et al., 1995; Montaggioni and Faure, 1997; Hopley and Davies, submitted for publication). Furthermore, the data presented herein indicate that the rates of vertical accretion of

Table 3  
Summary of vertical accumulation rates of framework and detrital facies, Holocene, Indo-Pacific coral reefs

Facies	Reef type	Reef pile zone	Water-energy	Rates (mm year <sup>-1</sup> )			Source
				Min	Max	Mean	
<i>Framework</i>							
Coralline algal	atoll	crest	high	0.1	2	1	Kayanne, 1992
	barrier, platform	crest, pavement	high	≤1	3	2	Davies and Hopley, 1983; Kayanne, 1992; Collins et al., 1993
	fringing	pavement	high/medium	≤1	3	2	Easton and Olson, 1976; Grigg, 1998; Engels et al., 2004
Robust-branching coral	barrier	outer flat	high	≤1	14	8	Camoin et al., 1997, 2004; Montaggioni et al., 1997a,b; Cabioch et al., 1999a; Webster, 1999; Kayanne et al., 2002
	platform fringing	outer flat outer flat foreereef	high high/medium	1 ≤1	12 15	6 8	Davies and Hopley, 1983; Collins et al., 1993; Glynn and Macintyre, 1987; Takahashi et al., 1988; Kayanne et al., 1993; Cortés et al., 1994; Montaggioni and Faure, 1997; Cabioch et al., 1998; Cabioch, 2003; Engels et al., 2004; Yamano et al., submitted for publication
Domal coral	atoll	inner flat	medium	NA	8	4.5	Marshall and Jacobsen, 1985; Ohde et al., 2002
	barrier, platform	outer/inner flats	high/medium	2	>13	4	Davies and Hopley, 1983; Montaggioni et al., 1997a,b; Cabioch et al., 1999a; Webster, 1999; Kayanne et al., 2002
	ocean-facing fr	outer/inner	flats high/medium	0.6	>16	5	Yonekura et al., 1988; Cortés et al., 1994; Cabioch et al., 1995; Kennedy and Woodroffe, 2000; Kayanne et al., 1993; Cabioch et al., 1998; Cabioch, 2003; Yamano et al., submitted for publication; Collins et al., 2003; Grigg, 1998; Grossman and Fletcher, 2004; Engels et al., 2004
	patch reef	lagoon	low	1	10	4	Marshall and Davies, 1982; Davies and Hopley, 1983; Kayanne et al., 2002
	inshore fringing	reef flat	low	1	10	4	Hopley et al., 1983; Davies and Hopley, 1983; Yamano et al., 2001a,b, 2003; Kleypas, 1996



Tabular-branching coral	barrier, platform	outer flat	high	3	30	7.5	Marshall and Davies, 1982; Davies and Hopley, 1983; Montaggioni, 1988; Montaggioni et al., 1997a,b; Cabioch et al., 1999a; Kayanne et al., 2002
	platform	inner flat/leeward	medium	3	9	5	Marshall and Davies, 1982; Davies and Hopley, 1983
	ocean-facing fr	outer/inner flats	high/medium	2	8.8	5	Kan et al., 1995; Kayanne et al., 1993; Yonekura et al., 1994; Montaggioni and Faure, 1997; Yamano et al., 2001a,b, submitted for publication; Kennedy and Woodroffe, 2002
Arborescent coral	inshore fringing	outer/inner flats	low	1	3	2	Davies and Hopley, 1983; Hopley et al., 1983
	barrier, platform	inner flat	medium	2	30	10	Davies and Hopley, 1983; Webster, 1999; Kayanne et al., 2002
	patch reef	lagoon	low	2	10	6	Marshall and Davies, 1982; Kayanne et al., 2002
	ocean-facing fr	inner flat	medium	1	25	9	Montaggioni, 1977; Takahashi et al., 1988; Collins et al., 1993; Yamano et al., 2001a,b, submitted for publication; Engels et al., 2004
Foliaceous coral	inshore fringing	outer/inner flats	low	1	5.5	3	Hopley et al., 1983; Davies and Hopley, 1983; Kleypas, 1996
	fringing	forereef/backreef	low	<1	6	3	Takahashi et al., 1988; Montaggioni and Faure, 1997
<i>Detrital</i> Rubble	atoll	inner flat/shallower lagoon	medium/low	<1	NA	3	Marshall and Jacobsen, 1985; Szabo et al., 1985; Ebrén, 1996
	barrier, platform	outer flat	high	3	18	10	Davies and Hopley, 1983; Woodroffe et al., 2000; Ohde et al., 2002
	barrier, platform	inner flat/backreef	medium/low	1	>10	5	Davies and Hopley, 1983; Cabioch et al., 1999a; Woodroffe et al., 2000; Kayanne et al., 2002; Montaggioni, unpublished
	ocean-facing fr	fore reef/outer flat	high	1.5	18	10	Yamano et al., submitted for publication; Engels et al., 2004
	ocean-facing fr	inner flat/backreef	medium/low	1	21	10	Glynn and Macintyre, 1987; Montaggioni and Faure, 1997; Yamano et al., submitted for publication; Kennedy and Woodroffe, 2000
	inshore fringing	flat/backreef	low	1	>20	8	Davies and Hopley, 1983; Hopley et al., 1983; Johnson and Risk, 1987; Kleypas, 1996; Zinke et al., 2003a,b

(continued on next page)

Table 3 (continued)

Facies	Reef type	Reef pile zone	Water-energy	Rates (mm year <sup>-1</sup> )			Source
				Min	Max	Mean	
Carbonate sand	atoll	inner flat/shallower lagoon	medium	0.2	1	>1	Tracey and Ladd, 1974; Smithers et al., 1992
	barrier, platform	outer flat	high	2	>20	10	Marshall and Davies, 1982; Davies and Hopley, 1983
	barrier, platform	inner flat	medium	<1	>10	4	Davies and Hopley, 1983; Woodroffe et al., 2000, 2004
	barrier, platform	shallower lagoon	low	<1	4	2	Marshall and Davies, 1982; Davies and Hopley, 1983; Smith et al., 1998
	ocean-facing fr inshore reefs	forereef flat/backreef	medium low	NA 0.1	NA 12	1.6 3	Cortés et al., 1994; Davies and Hopley, 1983; Hopley et al., 1983; Kleypas, 1996; Zinke et al., 2001
Mixed rubble and sand	barrier, platform	flat/shallower lagoon	high/medium	<1	12	6	Davies and Hopley, 1983; Smith et al., 1998; Woodroffe et al., 2000; Kayanne et al., 2002
	ocean-facing fr	flat/backreef	high/medium	1	>30	13	Cabioch et al., 1995; Yamano et al., 2001a,b, submitted for publication
	inshore fringing	flat/backreef	low	1	>40	15	Davies and Hopley, 1983; Montaggioni, 1988; Kleypas, 1996; Zinke et al., 2001
Carbonate mud	atoll	lagoon	low	NA	NA	1	Pirazzoli and Montaggioni, 1986
	barrier, platform	lagoon	low	0.3	2.1	1	Smith et al., 1998; Zinke et al., 2003a,b
	fringing	backreef	low	<1	5	2	Johnson and Risk, 1987; Kennedy and Woodroffe, 2000

fr=fringing.

Table 4  
Summary of vertical accumulation rates of backreef (lagoon) deposits, Indo-Pacific reef systems

Region	Site location	Reef type	Zone	Accumulation rate (mm year <sup>-1</sup> ) <sup>a</sup>			Source
				Min	Max	Mean	
Indian Ocean	Mauritius	fr	moat	1.3	3.7	2	Montaggioni and Faure, 1997
	Mayotte	fr	moat	NA	NA	4.63	Zinke et al., 2003a,b
	Mayotte	br	central deeper lagoon	0.2	4.39	2	Zinke et al., 2001
	Mayotte	br	proximal lagoon	0.1	1.9	0.5	Zinke et al., 2001
	Cocos-keeling	atl	shallower lagoon	0.25	1	0.75	Smithers et al., 1992
Pacific Ocean							
Ryukyus	Kabira	fr	moat	1	3	1.3	Yamano et al., 2001a,b
Great Barrier Reef	Stanley	pr	shallower lagoon	2.6	5	3	Davies et al., 1985
	One Tree	pr	shallower lagoon	0.5	1.7	1	Davies, 1983
	Heron	pr	central lagoon	0.7	2.1	1.4	Smith et al., 1998
	Boulder	pr	shallower lagoon	NA	NA	1.9	Webster, 1999
	Davies	pr	shallower lagoon	1.4	3.4	2	Tudhope, 1989
Tasman Sea	Lord Howe	fr	moat	1	11	5.3	Kennedy and Woodroffe, 2000
New Caledonia	South–West area	br	central lagoon	0.9	3.2	2	Ambatsian et al., 1997
Chesterfield island	Mouillage	atl	shallower lagoon	1.5	4.7	4.3	Degauge-Michalski, 1993
Marshall island	Majuro	atl	deeper lagoon	NA	NA	0.65	Yamano et al., 2002
Hawaiian island	Oahu	fr	moat	0.2	4	2	Easton and Olson, 1976
Cook island	Pukapuka	atl	shallower lagoon	1	11.7	2.8	Gray et al., 1992
	Rakahanga	atl	shallower lagoon	0.6	3.7	2	Gray et al., 1992
	Aitutaki	br	shallower lagoon	0.8	5.3	1.6	Gray (pers. com.)
Society island	Tahiti	br	shallower lagoon	1.2	15	5.7	Cabioch et al., 1999a
	Moorea	br	backreef flat	NA	NA	4.3	Montaggioni, 1988
	Huahine	br	shallower lagoon	NA	NA	6.9	Montaggioni, unpublished
Tuamotu island	Mataiva	atl	shallower lagoon	NA	NA	1	Pirazzoli and Montaggioni, 1986

Moat=backreef area of fringing reefs not exceeding 2 m deep; shallower lagoon=area of barrier reef and atoll systems close to the innermost parts of reef margins, not exceeding 10 m deep; deeper lagoon=area of barrier reef and atoll systems at depths greater than 10 m; proximal lagoon=area close to the inner-shelf fringing reefs; central lagoon=area situated at a distance from adjacent reef margins.

<sup>a</sup> The rates of vertical accumulation were estimated from the thickness of the deposits, measured perpendicular to the surface of the backreef floor.

coral assemblages do not appear to be controlled directly by the coral growth habits. However, although there is no clear variation in growth rates between branching and domal frameworks, the higher modal rates are recorded generally in cores containing a greater proportion of branching forms. Domal frameworks are normally characterized by growth rates of less than 7 mm year<sup>-1</sup> whereas those of branching habits are typically higher.

Abrupt variations in framework aggradation rates within a given cored sequence may relate to changes in the composition of the coral assemblage. This results from the response of the upward-growing reef pile to changes in environmental conditions. It is expressed in the replacement of a given coral community by one better adapted to the new conditions. For example, in Koror island (Palau), the

decrease in aggradation rates recorded from the outer barrier pile corresponds to the change in facies from the relatively loose structure of arborescent acroporids (30 mm year<sup>-1</sup>) to the rigid, robust acroporid framework (less than 3 mm year<sup>-1</sup>) as the reef top approached sea level within the 5-m depth interval (Kayanne et al., 2002). Similarly, beneath the leeward coral flat of the outer shelf reef Ribbon 5 (Great Barrier Reef), the increase of vertical accumulation rates upward (3.3 to 12 mm year<sup>-1</sup>) coincides with a change from domal to mixed domal/arborescent frameworks (Webster, 1999), as the result of the enhanced protecting effect by aggradation of the windward margin through time.

In detritus-dominated reef sections, the total variation in vertical accumulation rates is within the range of 0.2 to about 40 mm year<sup>-1</sup> (Table 3). The

Table 5  
Summary of vertical accretion rates of exposed reef-margin piles, Indo-Pacific province

Region	Site location	Reef type	Reef-margin zone	Vertical expansion rate (mm year <sup>-1</sup> ) <sup>a</sup>			Source	
				Min	Max	Mean		
Western Indian								
Madagascar	Toliara	br	reef flat	1.9	5.4	2.3	Camoin et al., 2003	
Mayotte		br	outer reef flat	4.3	8.3	6	Camoin et al., 1997	
Seychelles island	Mahé	fr	outer reef flat	2.3	8.8	3.5	Braithwaite et al., 2000	
Eastern Indian		Ningaloo	fr	upper fore reef	NA	NA	1.5	Collins et al., 2003
	Abrolhos	pr	outer reef flat	<1	7	3	Collins et al., 1993; Eisenhauer et al., 1993	
Western Pacific								
Ryukyus		Ishikagi-Jima						
	Kabira	fr	reef flat	0.2	5.8	2.9	Yamano et al., 2001a,b	
	Yoron-tou	fr	reef crest	1	3.2	2	Yonekura et al., 1994	
	Yoron-tou	fr	upper fore reef	5	8	12	Yonekura et al., 1994	
	Kuma-Jima	fr	reef flat	0.2	4.5	2	Takahashi et al., 1988	
	Kikai-Jima	fr	reef flat	2.2	5.8	3	Webster et al., 1998	
Great Barrier Reef								
Australia		Yonge	lbr	inner reef flat	1.8	20	6.5	Laurenti, 1995
	Ribbon 5	lbr	inner reef flat	1.3	14	7	Davies et al., 1985	
	One Tree	pr	outer reef flat	0.6	12	4	Marshall and Davies, 1982	
	Boulder	pr	outer reef flat	NA	NA	5.5	Davies et al., 1985	
	Stanley	pr	reef flat	>5	10.2	8	Davies et al., 1985	
Torres Straits		Yam	pr	reef flat	1.9	6.1	4	Woodroffe et al., 2000
	Warraber	pr	reef flat	1	9.8	7.3	Woodroffe et al., 2000	
Chesterfield island	Bellona	atl	reef flat	0.8	5	3	Degaugue-Michalski, 1993	
New Caledonia	Mamié	fr	reef crest	1.2	2.1	1.6	Cabioch et al., 1995	
Papua New Guinea	Huon	fr	reef crest	4.6	14	7	Ota and Chappell, 1999	
Vanuatu island	Espiritu Santo	fr	reef flat	1	5.8	2.5	Cabioch et al., 2003a,b; Cabioch, 2003	
Mariana island	Rota	fr	reef crest	0.6	14	5.7	Kayanne et al., 1993	
	Guam	fr	reef crest	0.9	6.9	3	Kayanne et al., 1993	
Marshall island	Enewetak	atl	reef flat	0.3	10	5	Tracey and Ladd, 1974; Szabo et al., 1985	
Kiribati island	Tarawa	atl	reef flat	1	8.2	5.4	Marshall and Jacobsen, 1985	
Tuvalu island	Funafuti	atl	reef flat	<1	4.6	3.2	Ohde et al., 2002	
Central Pacific								
Hawaiian island		Oahu (Sunset)	fr	fore reef	0.17	0.5	0.25	Grigg, 1998
	Oahu (Kailua)	fr	reef flat	0.25	6	2.5	Harney and Fletcher, 2003; Grossman and Fletcher, 2004	
	Oahu (Hanauma)	fr	outer reef flat	1	3	2	Easton and Olson, 1976	
	Molokai (Hale O Lono)	fr	fore-reef	2	23	7.5	Engels et al., 2004	
	Molokai (Hikauhi)	fr	fore reef	2.1	5.2	4	Engels et al., 2004	
Cook island	Mangaia	fr	reef flat	1	5	3.5	Yonekura et al., 1988	
Society island		Tahiti	br	reef crest	0.9	20.6	6.5	Montaggioni et al., 1997a,b
	Moorea	br	outer reef flat	3.3	6.7	5	Montaggioni, 1988	
Tuamotu island	Mururoa	atl	reef flat	1	2	1.8	Ebren, 1996	

Exposed reef-margin environments refer to fore reefs, reef crests and reef flats facing to ocean and subjected to prevailing winds and currents, i. e. situated in high fairweather water-energy settings (irrespective of the possible occurrence of storms).

Reef type: fr=fringing reef; br=barrier reef; lbr=linear barrier reef; pr=platform reef; atl=atoll.

<sup>a</sup> The rates of vertical expansion were estimated from the distance separating the starting (basal) accretion centre from the surface of the reef margin. The distance was measured perpendicularly to the reef surface.

Table 6

Summary of vertical accretion rates of semi-exposed to sheltered reef-margin piles, Indo-Pacific province

Region	Site location	Reef type	Reef-margin zone	Water-energy	Vertical expansion rate (mm year <sup>-1</sup> )			Source
					Min	Max	Mean	
Western Indian	Mayotte	ifr	outer flat	low	1	1.3	0.8	Zinke et al., 2003a
	Réunion	fr	inner flat	medium	1	6	3	Montaggioni, 1977; Camoin et al., 1997
	Mauritius	fr	outer flat	medium	1	4.7	3	Camoin et al., 1997; Montaggioni and Faure, 1997
Eastern Indian	Thailand, Pukhet	ifr	outer flat	low	20	30	25	Tudhope and Scoffin, 1994
	Abrolhos	pr	leeward flat	low	3.3	10.2	7	Collins et al., 1993; Eisenhauer et al., 1993
Western Pacific								
Ryukyus	Iki	fr	reef flat	low	NA	NA	2.3	Yamano et al., submitted for publication
	Okierabu-Jima	fr	reef flat	medium	0.3	8.8	6	Kan et al., 1995
	Ishigaki-Jima							
	Tonoshiro	fr	reef flat	medium	1.7	17.5	6.4	Yamano et al., submitted for publication
	Kabira	fr	reef flat	high	1.3	6.4	3	Yamano et al., 2001a,b
	Minna-Jima	fr	reef flat	medium	1.3	>3	2	Kan and Hori, 1993; Yamano et al., submitted for publication
	Tonaki-Jima	fr	reef flat	medium	2.8	9.2	5.9	Yamano et al., submitted for publication
	Kuma-Jima	fr	reef flat	medium	0.9	4	6.8	Takahashi et al., 1985
	Okinawa-Hontou	fr	reef flat	medium	<1	12.7	7.7	Yamano et al., submitted for publication
	Sekisei	fr	reef flat	medium	2.4	11.1	7	Yamano et al., submitted for publication
Palau island	Koror	br	reef crest	medium	2.2	30	6	Kayanne et al., 2002
		fr	reef flat	low	4	15	7	Kayanne et al., 2002
Great Barrier Reef, Australia	One Tree	pr	leeward flat	medium	1	15	6	Marshall and Davies, 1982
	Stanley	pr	leeward flat	medium	2.6	12	7	Davies et al., 1985
	Cape Tribulation	ifr	reef flat	low	3.5	5.1	4	Partain and Hopley, 1989
	Fantome	ifr	reef flat	low	2	23	7	Johnson and Risk, 1987
	Hayman	fr	reef flat	low	3	5	3.5	Hopley et al., 1983
	Orpheus	fr	reef flat	low	1.3	6.8	4.7	Hopley et al., 1983
	Rattlesnake	fr	reef flat	low	2	6.7	4	Hopley et al., 1983
	Penrith	fr	reef flat	low	3.6	67	10.3	Kleypas, 1996
	Cockermouth	fr	reef flat	low	<1	16	4	Kleypas, 1996
	High Peak	fr	reef flat	low	1.3	9.2	3.6	Kleypas, 1996
Tasman Sea	Lord Howe	fr	reef crest	medium	2.1	4.3	3	Kennedy and Woodroffe, 2000
New Caledonia	East coast	fr	reef flat	medium	1.2	12.5	7.3	Cabioch et al., 1995
	West coast	fr	reef flat	low	0.7	>30	5	Cabioch et al., 1995
	Tenia islet	br	inner flat	medium	–	–	1.2	Coudray, 1976
Central Pacific								
Society island	Tahiti	ifr	reef flat	low	NA	NA	36	Montaggioni, 1988
	Moorea	ifr	reef flat	low	1.6	>30	>10	Montaggioni, 1988
Eastern Pacific								
Costa Rica	Punta island	fr	forereef	medium	0.9	8.3	3.8	Cortés et al., 1994
	Punta island	fr	reef flat	medium	1.6	2.4	1.9	Cortés et al., 1994
Panama	Uva island	fr	reef flat	medium	1.1	4.8	3	Macintyre and Glynn, 1976
	Secas study	fr	forereef	medium	0.5	20.8	7.4	Macintyre and Glynn, 1976
	Saboga island	fr	reef flat	medium	1.2	1.4	1.3	Macintyre and Glynn, 1976

Semi-exposed to sheltered reef-margins zones refer to reef flats facing the ocean, but not directly subjected to breaking waves (i.e. the innermost parts or the leeward side of open reef margins protected from prevailing currents are categorized as medium-energy reef environments) and to reef fronts, reef crests and reef flats not facing the ocean and situated in inner-shelf settings (inshore reefs).

Reef type: fr=fringing reef; ifr=inshore fringing reef; br=barrier reef; lbr=linear barrier reef; pr=platform reef; atl=atoll.



Table 7  
Summary of lateral expansion rates<sup>a</sup> of Indo-Pacific coral reefs

Region	Site location	Reef type	Reef zone	Water-energy conditions	Rates (mm year <sup>-1</sup> )			Source	
					Min	Max	Mean		
Indian ocean	Mauritius	fr	outer margin	medium	30	>300	80	Montaggioni and Faure, 1997	
	Abrolhos	pr	outer margin	high	NA	NA	300	Eisenhauer et al., 1993	
	Pukhet	fr	outer margin	low	17	120	80	Tudhope and Scoffin, 1994	
Pacific ocean	Ryukyus	Ishikagi-Jima							
		Kabira	fr	windward margin	high	34	149	88	Yamano et al., 2001a, 2003
		Tonoshiro	fr	leeward margin	medium	24	76	42	Yamano et al., 2003, submitted for publication
		Iki	fr	outer margin	low	NA	NA	5	Yamano et al., 2001b
		Kuma-Jima	fr	outer margin	high	NA	NA	50	Takahashi et al., 1988
		Okierabu	fr	outer margin	medium	63	100	78	Yamano et al., submitted for publication
		Minna	fr	outer margin	medium	29	45	37	Kan and Hori, 1993; Yamano et al., 2003
Great Barrier Reef	One Tree	pr	windward margin	high	NA	NA	50	Marshall and Davies, 1982	
	One Tree	pr	backreef sandy wedge	low	NA	NA	40	Davies, 1983	
	Fantome	fr	outer margin	low	NA	NA	80	Johnson and Risk, 1987	
	Magnetic	fr	reef flat	low	NA	NA	84	Chappell et al., 1983	
	Hayman	fr	outer margin	low	9	124	31	Kan et al., 1997a,b,c; Yamano et al., 2003	
	Orpheus	fr	outer margin	low	NA	100	46	Hopley et al., 1983	
	High Peak	fr	outer margin	low	8	40	25	Kleypas, 1996	
Cockermouth	fr	outer margin	low	NA	NA	84	Kleypas, 1996; Yamano et al., 2003		
Torres Straits	Yam	pr	reef flat	high	NA	NA	84	Woodroffe et al., 2000	
New Caledonia	Ricaudy	fr	outer margin	low	NA	NA	15	Cabioch, 1988	
Mariana island	Rota	fr	outer margin	high	5	21	10	Kayanne et al., 1993	
	Guam	fr	outer margin	high	NA	NA	40	Kayanne et al., 1993	
Palau island	Koror	br	outer margin	high	NA	NA	10	Kayanne et al., 2002	
Hawaiian island	Oahu	fr	outer margin	medium	15	30	22	Easton and Olson, 1976	
	Molokai (Hale O Lono)	fr	fore-reef	high	100	220	150	Engels et al., 2004	
Tuamotu island	Mururoa	atl	outer margin	high	NA	NA	24	Ebren, 1996	
Costa Rica	Punta island	fr	fore-reef	medium	NA	NA	52	Cortés et al., 1994	

Reef type: fr=fringing reef; br=barrier reef; pr=platform reef; atl=atoll.

<sup>a</sup> The rates of lateral expansion were estimated from the distance separating the position of the identified starting accumulation centre from that of its modern counterpart within a considered reef pile. The distance was measured parallel to reef-top surface.

present study largely confirms the pattern of detrital sedimentation established by Davies and Hopley (1983) in the Great Barrier Reef. Three accumulation-rate populations can be identified throughout the Indo-Pacific reef province, related to an increasing hydrodynamic-energy gradient: (1) lower modal rates of 1–3 mm year<sup>-1</sup> (maximum rates: 11 mm year<sup>-1</sup>), representing lagoonal sedimentation of mud-dominated material; (2) modes of 4 to 8 mm year<sup>-1</sup>, reflecting steady accumulation of sand to rubble

within reef flat and backreef environments under fairweather conditions; (3) higher modal rates (up to 10 mm year<sup>-1</sup>), implying rapid deposition of sand to rubble, presumably controlled by winter storms associated with trade-winds or by hurricane-generated waves. The highest rates of detrital accumulation are generally recorded from fringing and narrow platform reefs. In such sites, rates of deposition generated by low-frequency, high-energy events are at least 2 orders greater than those from large shelf-reef

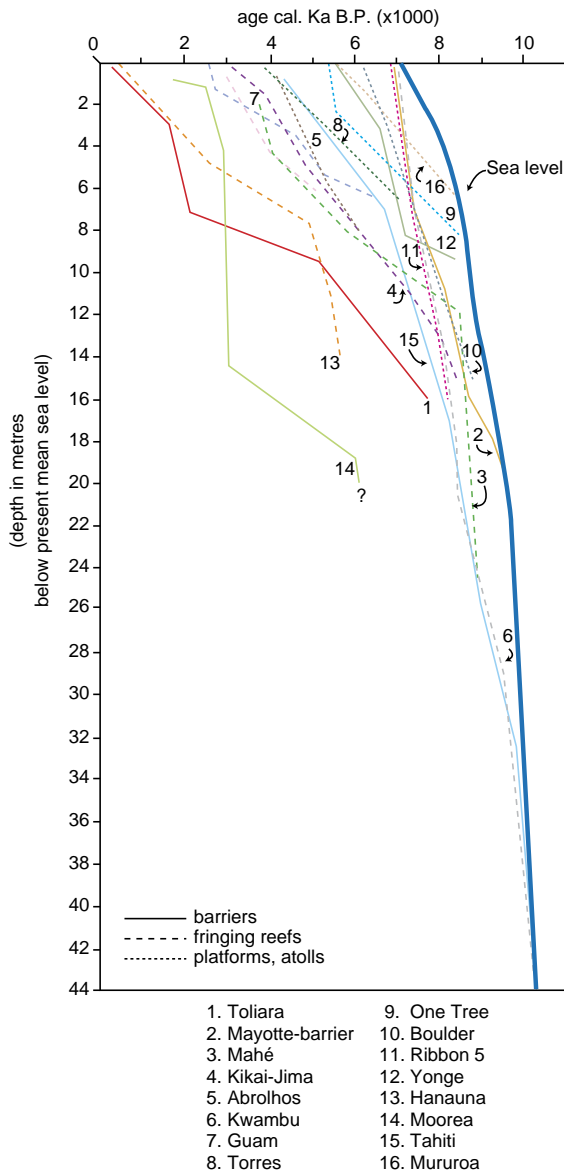


Fig. 9. Curves of vertical reef accretion from exposed reef crests/flats of barriers, fringing tracts, platforms and atolls. (see Fig. 2 for the list of references from which the curves are extracted). The selected reef piles are regarded as deposited through aggradation mainly. Sea level curve from the western/central Pacific.

systems. This may be due to the rapid infilling of the restricted and shallower inner reef basins, in contradistinction with large, widely open reef systems that, in addition, are washed away by strong currents. Rates of detrital deposition also vary in backreef areas with

depth (Table 4). Higher rates have been reported from moats and shallower lagoonal parts (1 to 6.9 mm year<sup>-1</sup> on average); lower rates relate to the deeper parts of large barrier reefs and atolls (0.65 to 2 mm year<sup>-1</sup>). These differences in accumulation rates have to be ascribed probably to the nature of depositional processes; whereas the proximal parts of backreef zones are supplied mainly with allochthonous material derived from adjacent reef margins, sedimentation in most deep lagoons results from in-situ carbonate production (Adjas et al., 1990; Yamano et al., 2002).

The accretional efficiency of reef margins seems to depend upon exposure to wave agitation. In exposed settings, the mean aggradation rates range between 1.5 and 12 mm year<sup>-1</sup>, with a mode of 5 mm year<sup>-1</sup> (Table 5). In semi-exposed to sheltered reef margins, vertical accretion rates average 1 to 25 mm year<sup>-1</sup>, with a mode of 9 mm year<sup>-1</sup> (Table 6). By contrast, in high-energy reef margins, lateral expansion rates vary from 24 to 300 mm year<sup>-1</sup>, with a mode of 90 mm year<sup>-1</sup>. Medium- to low-energy margins have developed laterally at rates averaging 15 to 84 mm year<sup>-1</sup>, with modal rates of about 50 mm year<sup>-1</sup> (Table 7; also see Table 1 in Yamano et al., 2003). Comparison of these data indicate that the rates of aggradation and lateral accretion (progradation mainly) appear to correlate negatively. This pattern probably results from changes in environmental conditions during reef development. Before sea level has stabilized, the dominating accretional process in most Indo-Pacific reefs has been aggradation tending to rapidly balance sea-level rise. Two reasons may explain why reef margins have developed vertically slower under higher energy conditions. Firstly, framework in these sites are made up mainly of slower-growing coral forms including robust branches and domes, compared to faster growing arborescent and tabular forms found mainly in medium- to low-energy sites; the great part of detritus has been washed away and accumulate in innermost flat and backreef areas. Secondly, once reef tops have been within depths less than about 5–6 m, high-wave energy may have depressed framework development (Grigg, 1998; Grossman and Fletcher, 2004). After sea-level stabilization and total infilling of vertical accommodation space, most reef margins have prograded seawards. The more extensive development of reef margins in higher-energy environments may be related to coral

biology and ecology, i.e. larval settlement, coral metabolism, food and nutrient supply, sediment removal (Yamano et al., 2003). All of these factors are encouraged in high-energy wind-wave fields, whereas lower-energy reef margins and, particularly, from inner-shelf settings, are usually subject to extremes in salinity and turbidity.

Rates of accretion also vary with time. As firstly shown by Davies and Marshall (1980), Holocene reef accumulation rates may have changed approximately in the pattern of a sigmoidal curve, with slow rates during earlier coral colonization of substrates and later when reef surface approaching sea level. The lower-most part of the sigmoidal curve expresses rates less than  $2 \text{ mm year}^{-1}$ , regarded as relating to inimical ambient conditions during reef initiation. The upper-most part reflects reduction in accretion rates (less than  $4 \text{ mm year}^{-1}$ ) as coral framework and algal crusts have provided the final capping to the reef in the form of reef crest (Kayanne, 1992). The bulk part of the curve expresses maximum rates of growth of between 5 to up to  $10 \text{ mm year}^{-1}$  for core sections accumulated under optimal conditions. It seems that the sigmoidal shape of reef growth curves typifies the catch-up growth mode. From keep-up reef sequences, the slow start-up growth event generally is missing, since aggradation maintained pace with rising sea level as soon as the substrates were flooded. The highest rates of vertical accretion usually coincide with one of two events: (1) reef initiation, when the development of pioneer communities tends to keep pace with rising sea level; (2) main growth phase, when, after a jump in sea level, reefs increased aggradation rates to escape drowning. Variations in growth rates therefore appear to be linked in part to growth styles.

The rates and modes of reef growth vary greatly within the same reef system, according to the considered zone (Davies et al., 1985; Montaggioni, 1988). For example, in the barrier reef complexes of Tahiti (Montaggioni, 1988; Cabioch et al., 1999a) and Palau (Kayanne et al., 2002), the keep-up mode operated in the windward, outer margins at mean rates of  $6 \text{ mm year}^{-1}$ , whereas the catch-up style controlled aggradation at rates of  $3\text{--}4 \text{ mm year}^{-1}$  in the leeward, inner margins and in lagoonal patchreefs. Conversely, growth modes are independent of both reef facies, form and setting. Both catch-up and keep-up signatures can pertain to

framework as well as detritus, in ocean-facing or sheltered reef flats (Figs. 9 and 10). For example, at Tahiti, the windward barrier framework as well as detrital inner fringing reefs developed through a keep-up mode (Montaggioni, 1988). In Mauritius island, in the medium-energy fringing reef of La Pointe-au-Sable, only the catch-up growth signatures were identified across the entire reef system, from the reef crest to moat piles (Montaggioni and Faure, 1997). Furthermore, growth styles seem not to depend upon the depth from which individual reefs initiated within the photic zone. Catch-up accretion can have initiated at depths as great as 40 m (Montaggioni, 1988).

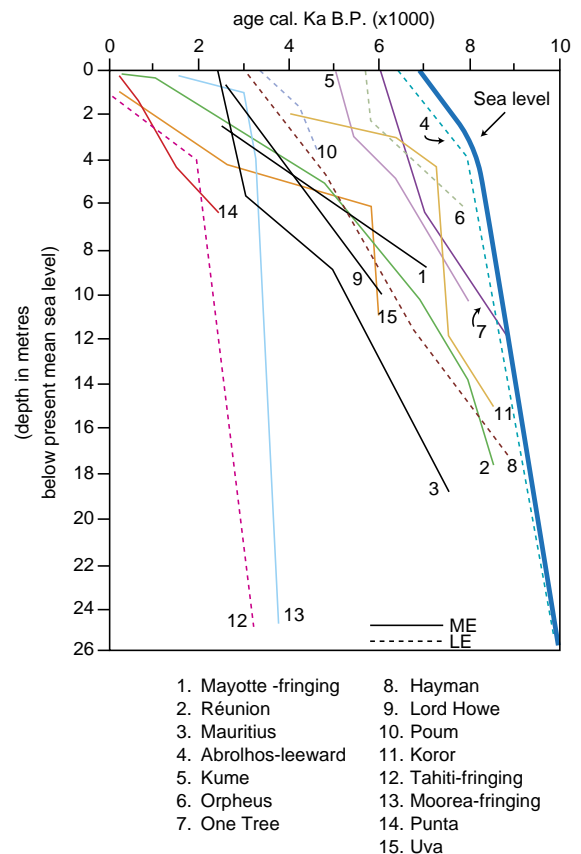


Fig. 10. Curves of vertical reef accretion from semi-exposed to sheltered reef crests/flats of fringing tracts and platforms (see Fig. 3 for the list of references from which the curves are extracted). The selected reef piles are regarded as deposited through aggradation mainly. ME=medium energy (semi-exposed); LE=low energy (sheltered).

Kayanne (1992) pointed out that a “keep-up” mode of reef development is more common in the Caribbean than in the Pacific, suggesting that it was encouraged by lower rates of sea-level rise in the Atlantic in the last 8 ka. This assertion is open to question, because few data are available from drilling high-energy outer reef zones. There is no clear evidence to suggest that the outer reef margins of Indo-Pacific reefs, dominantly consisting of branching coral framework, would not have been able to develop from a “keep-up” growth style, particularly in ocean-facing settings. Furthermore, Blanchon and Blakeway (2003) questioned the validity of the conceptual model of reef growth responses to sea-level rise. They attempted to demonstrate that the “catch-up” signatures may be artefacts of drilling and reefs may have grown more commonly through the “keep-up” mode than previously indicated. As mentioned above, the responses of reefs to sea-level rise are well known to vary greatly within the same system. An entire reef system can rarely be categorized as a “keep-up” or “catch-up” reef. It is clear that any curve of vertical accretion obtained from a single core strictly records the response of the coral assemblages at the coring site and is not representative of the overall development history of the reef.

Davies and Hopley (1983) noted that there is little variation in growth rates attributable to latitudinal differences. Neither framework growth nor detrital sedimentation rates are suppressed with increasing latitude. For instance, in the highest-latitude reefs, such as those of Iki Island (Japan), 33°48' north, aggradation rates of coral framework were up to 8 mm year<sup>-1</sup> (Yamano et al., 2001b). On Middleton and Elizabeth Reefs and Lord Howe Island, situated in the Tasman Sea (between 29°7' and 33°30' south), lagoonal deposition operated at mean rates of 2 to 5 mm year<sup>-1</sup> during the mid-Holocene. Such rates are as high as any recorded from tropical reef frameworks and greater than those calculated for many low-latitude lagoons. Kennedy and Woodroffe (2000) and Woodroffe et al. (2004) concluded that reefal carbonate sedimentation does not seem to be reduced in areas close to the latitudinal limits of reef growth, in spite of the slight decrease in the rate of modern sedimentation. However, and by contrast, in the Hawaiian archipelago stretching from about 19°5' to 28°3' north, rates of reef

growth have been shown to decrease as a linear function of increasing latitude, ranging between 11 and 0.2 mm year<sup>-1</sup>. During the Late Holocene, reef growth on Kure and Midway atolls, at the NW end of the chain, was not able to track rising sea level (Grigg, 1982).

In summary, based on maximum rates of vertical accretion, the Indo-Pacific reefs can be classified into three types:

- (1) fast-growing reefs, with rates up to 10 mm year<sup>-1</sup>. Such rates can be sustained for 3–5 ka and may reach up to 20 mm year<sup>-1</sup> for periods of about 500 years. They result commonly in expanded sequences up to 25 m thick.
- (2) moderate-growing reefs, with rates of 5–7 mm year<sup>-1</sup>, generally produce sequences 10–25 m thick.
- (3) slow-growing reefs, with rates of 1–4 mm year<sup>-1</sup> commonly form sections less than 10 m thick.

Vertical accretion potential can be converted to net calcification (Smith, 1983), a measure of carbonate production expressed in kg CaCO<sub>3</sub> m<sup>-2</sup> year<sup>-1</sup>, taking into account a value for the porosity of the original framework and detritus fabric (about 50%) and the density of an aragonite and calcite mixture (about 2.89 g cm<sup>-3</sup>). Thus, in these terms, fast-growing reefs release on average 10 kg CaCO<sub>3</sub> m<sup>-2</sup> year<sup>-1</sup>. This is close to values recorded from active modern reef crests (Kinsey, 1983). Using the same assumptions, moderate-growing reefs are typified by rates of 3–6 kg m<sup>-2</sup> year<sup>-1</sup> while rates for slow-growing range from 1–3 kg m<sup>-2</sup> year<sup>-1</sup>. As accumulation rates changed, so did carbonate production. Ryan et al. (2001) pointed out, from the study of the Holocene growth history of Wistari Reef (southern Great Barrier Reef), that production of the reef-top surface peaked at about 9 kg m<sup>-2</sup> year<sup>-1</sup> between 7 and 4 ka BP and has since decreased linearly to the present day. The implication is that mid-Holocene calcification rates must have been as high as 15 kg m<sup>-2</sup> year<sup>-1</sup>.

#### 3.4. Timing of reef initiation and development

For 23 ka, the sea-level history of the tropics has been punctuated by phases in which periods of reef initiation were followed by reef demise. Four gen-

erations of reefs have tentatively been recognized, locally separated by major non-constructural or reef drowning events (Fig. 11a,b).

### 3.4.1. Generation RGO

Estimates of sea level during the LGM interval are  $125 \pm 4$  m lower than now (Yokoyama et al., 2001; Peltier, 2002). There is still considerable uncertainty as to how coral growth responded to LGM sea surface temperatures (SSTs) (see Growth-controlling factors) and what the geographical distribution of reefs or at least of coral populations, might have been, principally because most of the indicators that are preserved are at present submerged. However, RGO bodies have been recorded in both oceans, from submersible diving and dredging on deep forereef slopes and drilling of tectonically emerged reefs.

On Mayotte (Comoros), as a result of the steepness of the foreslope, coral assemblages, dated at around 18.4 ka BP and at present at depths of about 150 m, developed only as thin veneers. They consist of in-situ shallow-water scleractinians (*A. gr. robusta*, and domal *Porites*) together with crusts of the coralline alga *H. onkodes*. Downslope, below 180–220 m present-day water depth, coral material, apparently derived from shallow-water, has been dated at 19.7–19.4 ka BP (Dullo et al., 1998). On the rapidly uplifting island of Espiritu Santo and nearby islets (Vanuatu), reef growth started at 24.3 to 23 ka BP. The

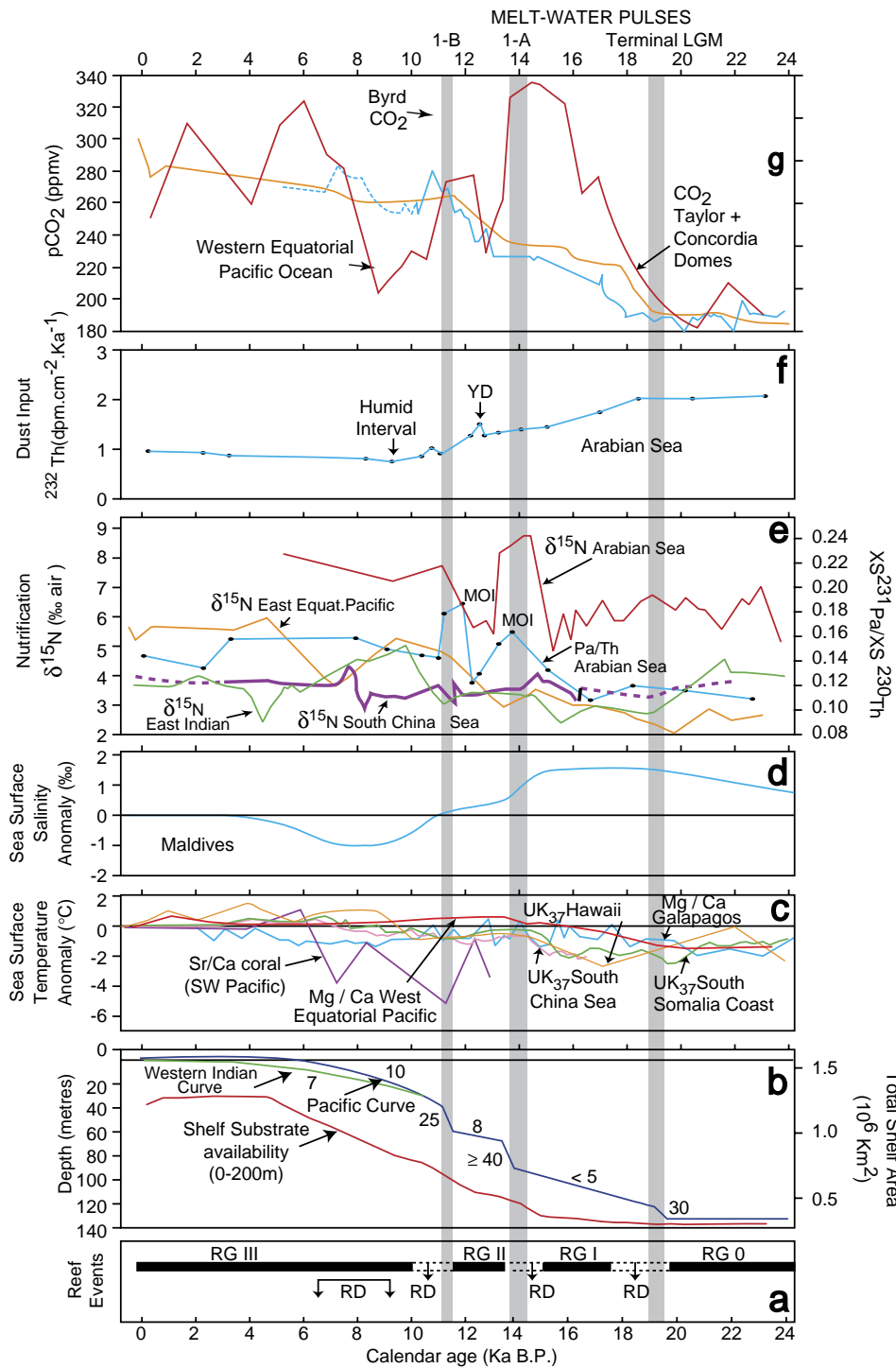
relevant communities, dominated by tabular acroporids (*A. gr. hyacinthus*), showed an upward accretion at up to  $1 \text{ mm year}^{-1}$ , indicating they were true reef communities rather than simple coral assemblages (Cabioch et al., 1998, 2003a). This LGM sequence is only 2–3 m thick and is found between about 20 and 60 m below present sea level. At Rendova (Solomon islands), drillholes penetrated corals of LGM age about 65 m below present sea level (Taylor et al., 2000). In the Marquesas archipelago (French Polynesia), samples of *Acropora* 20 ka BP old were dredged from reefal relicts on 125 to 100 m-deep submarine platforms (Cabioch et al., 2000, 2003b). A faunal-rich reef of probable glacial age was described from the insular shelf off the Miyako islands in the Ryukyu group (Yamano et al., 2001b). There is no sustained evidence of RGO features along the foreslopes of the Australian Great Barrier Reef (Carter and Johnson, 1986; Hopley et al., 1997).

The LGM terminated at around 19 ka BP with a rise in sea level that was sustained at a rate of about  $30 \text{ mm year}^{-1}$  for at least 500 years (Yokoyama et al., 2001). This terminal-LGM meltwater pulse (Lambeck et al., 2000) marked the rapid decay of the polar ice sheets. Presumably as a result of the limited tolerance of the main reef builders to the environmental conditions of the LGM, coral growth was unable to keep up with this rapid rise. For this reason, apart from sites subject to intense tectonic uplift, incipient

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Fig. 11. Summary of the major environmental events in the last 23 ka controlling reef growth in the Indo-Pacific province at global to regional scales (modified from Montaggioni, 2000). The vertical bands indicate the timing of the major meltwater pulses (Terminal-LGM MWP, MWP-1A and MWP 1-B) identified at 19, 14 and 11.5 ka BP, respectively. (a) Schematic reconstruction of reef evolution, with indication of reef growth phases (RGO, RGI, RGII, RGIII) and non-constructural or reef drowning (RD) events. (b) Last Glacial to deglacial sea-level curve showing rates of sea-level rise ( $5$  to about  $40 \text{ mm year}^{-1}$ ) at various stages of the rise (data from Fairbanks, 1989; Bard et al., 1990; Chappell and Polach, 1991; Bard et al., 1996; Hanebuth et al., 2000; Yokoyama et al., 2001; Weaver et al., 2003). Note that, from about 10 ka BP, the sea-level history in the western Indian Ocean and the Pacific, has varied significantly, due to differential redistribution of water masses (see Camoin et al., 1997 for discussion). Also shown is the curve of changes in shelf substrate availability (between 0- and 200-m depth) between the Last Glacial Maximum and the present (data from Kleypas, 1997). (c) Curves of sea surface temperatures: the SW Pacific curve is based on the Sr/Ca ratio measured from shallow-water corals colonies (Beck et al., 1997); the curves from South China, South Somalia and Hawaii are based on the analysis of alkenones in ocean-bottom sediments (data from Steinke et al., 2001; Sonzogni et al., 1998; Lee et al., 2001, respectively); the curves from the Galapagos region and the West Equatorial Pacific were established on the basis of Mg/Ca ratios in benthic foraminifers (data from Koutavas et al., 2002; Palmer and Pearson, 2003, respectively). (d) Curve of sea surface salinity from Maldives, Western Indian Ocean (data from Rostek et al., 1993). (e) Changes in nutrient utilization: distribution of bulk sediments  $\delta^{15}\text{N}$  values in the South China Sea (Kienast, 2000, modified by Steinke et al., 2004), in the East Equatorial Pacific (Farrell et al., 1995), in the East Indian Ocean (Muller and Opdyke, 2000) and in the Arabian Sea (Suthhof et al., 2001), and changes in paleoproductivity: the  $^{231}\text{Pa}/^{230}\text{Th}$  ratio expresses productivity as a function of intensity of upwelling-favourable monsoonal winds (MOI=monsoon intensification events) (data from Marcantonio et al., 2001). (f) Changes in dust supply: fluxes of  $^{232}\text{Th}$  in Arabian Sea sediments express changes in the accumulation rates of wind-blown dust from the Arabian Peninsula and Persian Gulf region (YD=Younger Dryas) (data from Marcantonio et al., 2001). (g) Changes in  $p\text{CO}_2$  from Antarctic Ice-cores: Byrd Station (Neftel et al., 1988; Blunier et al., 1997), Taylor Dome (Indermühle et al., 1999), and Dome Concordia (Monnin et al., 2001), and from surface waters in the Western Equatorial Pacific (Palmer and Pearson, 2003).





buildups formed before 19 ka BP were all drowned. Grigg and Epp (1989) suggested that, in the Pacific, the drowning of many atolls during the last deglaciation was controlled by the elevation of atoll summits relative to paleosea level at the end of the LGM. The rapid rise in sea level may have been responsible for the demise of about one-third of all Pacific atolls. Only those that were sufficiently elevated, now at depths more than 100 m below sea level, were able to track the rising sea level. A similar scenario was proposed by Purdy and Bertram (1993) for explaining the existence of drowned banks in the Maldives. Thus, most of the submerged coral banks in the North and Central Pacific (Hawaii, the Marshall islands and Tuamotu archipelago) and in the Indian Ocean (Laccadives, Chagos, and the Amirantes) may have been below the critical depth threshold.

#### 3.4.2. Generation RGI

Clark and Mix (2000) suggested that the rapid rise in sea level at 19 ka BP was associated with an ocean-cooling event. It is difficult to evaluate to what extent reef colonization may have been perturbed by the combination of these inimical factors. There is no evidence in the Solomon and Vanuatu islands, lying within the West Pacific Warm Pool and characterized by the warmest sea surface temperatures ( $\geq 28$  °C) in the open ocean, of prominent hiatuses in coral reef deposition from the LGM termination to about 15 ka BP (Cabioch et al., 1998, 2000, 2003a; Taylor et al., 2000). On Mayotte, reworked coralline debris at depths ranging from 180 to 285 m and dated at 19.1 to 18.0 ka BP was derived from a thriving shallow-water reef environment (Dullo et al., 1998). These findings indicate that in many areas, coral communities were flourishing during the 19–18 ka BP event. However, along many shelf margins, rapid flooding may have prevented the early establishment of significant coral faunas (Great Barrier Reef, New Caledonia). In these areas, there was a time lag in colonization that has lasted more than 5 ka after flooding (Harris and Davies, 1989; Cabioch et al., 1999c).

From about 18.5 to 15 ka BP, although there is a critical gap in the data between 17 and 15 ka BP, sea level is interpreted to have risen uniformly at a rate not exceeding  $5 \text{ mm year}^{-1}$  (Fleming et al., 1998), probably at about  $2.3 \text{ mm year}^{-1}$  (Weaver et al., 2003)

due to a minimal meltwater influx to the oceans (Lohmann and Schulz, 2000). Relicts of scleractinian assemblages as old as 18–17 ka BP have been sampled at a few sites. On Mayotte, displaced elements of near-surface coral communities (*A. gr. robusta* and *G. fascicularis*), dated at 18–16 ka BP, have been recovered between 118- and 160-m depth (Dullo et al., 1998). In the Central Great Barrier Reef, specimens of *Galaxea clavus* with an age of 17 ka BP have been found as re-deposited detritus 175 m deep (Veeh and Veevers, 1970). Carter and Johnson (1986) assumed that the onlap of the reef belt on the GBR shelf-edge at 75 m deep relates to a 15-ka-BP-old shoreline. Bard et al. (1992) obtained an age of 17.6 ka BP from shallow-water corals on the flanks of Mururoa atoll (central Pacific), and Rao et al. (2003) reported the occurrence of pinnacles and reef-like structures at depths of 105–110 m below present sea level on the outer continental shelf of western India. These consist of in situ *Porites* colonies ranging in age from 15.5 to 14 ka BP.

The interval from 14.7–14.3 ka BP is characterized by a massive input of meltwater, culminating near 14 ka BP (MWP-1A pulse of Fairbanks, 1989) related to the Bølling–Allerød warm period (Groottes and Stuiver, 1997; Lohmann and Schulz, 2000; Hanebuth et al., 2000; Kienast et al., 2003). This caused a rise in sea level of about 15 m in less than 500 years, and is well documented in the Pacific (Bard et al., 1996; Hanebuth et al., 2000). The rates of sea-level rise, variously estimated at up to  $40 \text{ mm year}^{-1}$  (Bard et al., 1990; Blanchon and Shaw, 1995; Weaver et al., 2003) exceeded the ability of vertical reef accretion to keep up and with the exception of areas where the transgression was compensated by uplift (Vanuatu: Cabioch et al., 1998) this period is marked by non-construction or reef drowning. There is little evidence of submerged reef tracts of Bølling age in the Indo-Pacific. Along the deep forereef of the Great Barrier Reef, reef-like structures have been found at present depths of about 90 m (Hopley, 1982; Harris and Davies, 1989; Hopley et al., 1997) that are thought to have drowned during the MWP-1A event. Similar features also occur at 90–100-m depth in the Comoro islands (Dullo et al., 1998), and at 150–160 m below present sea level on Hawaii (Moore et al., 1990). Webster et al. (2004) clearly identified 150-m-deep reefs dated at 15.2–14.7 ka BP off Hawaii and

interpreted as drowned by rapid sea-level rise associated with MWP-1A.

#### 3.4.3. Generation RGII

As the effects of meltwater discharge decreased, between about 13.8 and 11.5 ka BP, sea level in the Central Pacific rose at a rate averaging  $7.5 \text{ mm year}^{-1}$  (Bard et al., 1996), comparable to the rate in the Caribbean ( $8 \text{ mm year}^{-1}$ , Bard et al., 1990) for the same period. This eustatic rise accompanied a rapid warming (Gagan et al., 2000) that resulted in the initiation of a new reef generation (RGII). The latter is well documented in the western and central Pacific regions (Chappell and Polach, 1991; Montaggioni et al., 1997a; Cabioch et al., 1998, 1999a; Taylor et al., 2000). On the Huon Peninsula (New Guinea), from 13.2 to 11.8 ka BP, reef crest and upper forereef zones developed upward at an average rate of  $8.7 \text{ mm year}^{-1}$  to a maximum thickness of around 13 m. Growth of coral communities, dominated by branching acroporids (*A. gr. hyacinthus*, *A. palifera*), pocilloporids, and domal *Porites* (*P. lobata*), generally kept pace with rising sea level. In Tahiti, shelf inundation occurred around 14 ka BP and reef colonization probably started at the same time with a robust-branching *Acropora* framework.

Contemporaneous, submerged reef-like bodies have been described from various continental and island shelves. Along the western outer shelf margin of India, at depths of about 75 m, Rao et al. (2003) reported 4- to 14-m-high coralgall buildups, dated at about 14 ka BP. On Mayotte, coralgall mounds, dated at 13.6 to 12 ka BP in age, lie 55–90 m below present sea level. These consist of in-situ corals dominated by *Porites*, and *Pocillopora*, encrusted by coralline algae, vermetid gastropods and foraminifers (Dullo et al., 1998). Most reefs that developed within the 13.8–11.5 ka BP interval are at present encountered as relict features and terraces at depths of 50–90 m. Numerous undated reefs within the same depth range are found throughout the Indo-Pacific, with examples on Mauritius (Faure and Montaggioni, 1976); Madagascar (Pichon, 1978); Seychelles (Montaggioni, 2000); the Great Barrier Reef (Hopley, 1982; Carter and Johnson, 1986; Harris and Davies, 1989; Hopley et al., 1997); New Caledonia (Coudray, 1976); and the Marquesas (Cabioch et al., 2003b). This reef submergence event was probably triggered by meltwater

discharge at around 11.5 ka BP (MWP-1B pulse of Fairbanks, 1989). The resulting jump in sea level of significantly smaller magnitude than MWP-1A (Bard et al., 1996) caused a major break in reef development. However, locally reef margins were able to compensate for the abrupt increase in accommodation. For instance, in Vanuatu, this event seems to be verified by an abrupt change in the composition of coral communities, from a shallower-water, acroporid-dominated assemblage to a deeper, poritid-dominated one, that occurred between 12.6 and 11.3 ka BP (Cabioch et al., 2003a). On the subsiding island of Tahiti (Cabioch et al., 1999a) and along the rapidly uplifting, northeast coast of the Huon Peninsula (Ota and Chappell, 1999), reef crests grew vertically at rates of up to  $10 \text{ mm year}^{-1}$  between 11.5 and 11.3 ka BP, respectively, and were able to remain close to or to catch up with sea level.

#### 3.4.4. Generation RGIII

There was a prominent gap in reef growth at the end of the MWP-1B event. In the Indo-Pacific, sea-level rose at a moderate rate of about  $10 \text{ mm year}^{-1}$  from 11 to about 7–6.5 ka BP until it stabilized around its present position (Pirazzoli, 1996). Apart from reefs that escaped demise and continued to develop, no re-settlement of reef-building communities seems to have begun before 10 ka BP and most modern reefs can be related to this generation. Throughout the Indo-Pacific, they commenced growth within a relatively restricted period (Montaggioni, 1988) and initiation was bracketed between 10 and 7 ka BP. Generally, when sea level reached a position 30–40 m below the present datum, coral populations were able to colonize inner shelf substrates and reefs that were still active began to flourish in shallow coastal areas (Carter and Johnson, 1986). Recolonization may have been favoured by improved shelf-water quality during this period (Davies et al., 1985; Montaggioni, 1988, 2000). As a consequence, Late Holocene near-surface reef tracts range from 10 to 30 m thickness. Locally, however, re-settlement seems to have delayed and started more recently. In New Caledonia, Northern Australia and the eastern Pacific, growth pauses were common and reefs formed only thin veneers (1–6 m) with a relief locally inherited from non-carbonate substrates. These structures range in age from 5.6 ka BP to modern (Macintyre et al., 1992; Cortés et al.,

1994; Cabioch et al., 1995; Glynn and Ault, 2000; Smithers and Larcombe, 2003). A similar pattern is reported locally from the latitudinal limits of reef growth. In the Ryukyu Islands (Japan), some fringing reefs are probably younger than 4 ka BP (Yamano et al., 2001b). Such pauses or slowing of reef growth may have been induced in part by short periods of rapid sea-level rise between 9 and 7 ka BP. Blanchon et al. (2002) reported the possible occurrence of a rapid 6-m jump in sea level at around 7.5 ka BP from the study of a submerged reef at a depth of about 20 m in the Caribbean.

#### 4. Growth-controlling factors

##### 4.1. Sea level

Global variations in sea level related to the last deglaciation have determined, in conjunction with neotectonics and antecedent topography, the location and geometry of reefs. As emphasized by Hubbard (1988), variations in these factors have resulted in very different sea-level histories reflecting site-specific reef growth scenarios. Using a hydraulic model of the water exchange between the Red Sea and the world ocean, Siddall et al. (2003) found that sea-level changes of up to 35 m, at rates of up to 20 mm year<sup>-1</sup> have occurred during the last glacial cycle, coincident with abrupt climatic changes. More particularly, the abrupt changes in the rate of sea-level rise at 19, 14 and 11.5 ka BP, commonly caused reef drowning (Fig. 11a,b). This helps to explain why there are at least two to three generations of submerged tracts on many tropical shelf margins (Fig. 12): Madagascar (Jouanic, 1972), Mayotte (Dullo et al., 1998), New Caledonia (Coudray, 1976), the Great Barrier Reef (Carter and Johnson, 1986; Harris and Davies, 1989; Hopley et al., 1997), the Marquesas (Cabioch et al., 2003b). However, reef crests were able to keep pace with the rapidly rising sea level, forming expanded Postglacial sequences, in both subsiding (87 m thick on Tahiti) and rapidly uplifting areas (up to 60 m thick on the Huon Peninsula and Vanuatu). Two hypotheses could explain why some Indo-Pacific reefs have been able to compensate for some sharp and episodic variations in sea-level rise. Each jump in sea level may have been the sum of several low-amplitude,

high-frequency pulses, including brief still-stands; or, as suggested by Blanchon (1998), reefs survived the jumps whose magnitude was not sufficient to displace the corals at the reef crests out of their habitat depth interval (0–6 m). The MWP-1B and the suspected jump at around 7.5 ka BP had much smaller magnitude than MWP-1A and therefore apparently did not affect reef development in the same way. The reefs were able to recover leaving no resolvable framework record of the events (Blanchon et al., 2002).

In the Indo-Pacific, reef tops commonly reached the sea surface several millennia ago as sea level stabilized in the 7–3 ka BP period. This stillstand episode promoted forward progradation of reef margins locally extending 250 to 1000 m (Johnson and Risk, 1987; Tudhope and Scoffin, 1994; Yamano et al., 2001a, 2003) and the backward transport and deposition of detritus leading to the formation of extensive reef flats (Grossman and Fletcher, 2004). Middle to late Holocene exposed reef flats lying between about 0.5 and 3 m above present sea level are usual features in the tropical Pacific Ocean basin (Pirazzoli, 1991; Nunn, 1994; Grossman et al., 1998; Dickinson, 2001). When not overprinted by local uplift, thermal subsidence or lithospheric flexure, the exposure of RGIII reefs is thought to result from mid-Holocene eustatic rises in sea level, combined with late Holocene sea-level falls induced by the process of glacial isostatic adjustment (Mitrovica and Milne, 2002). This process, so-called “equatorial ocean syphoning” (Mitrovica and Peltier, 1991) relates to the migration of meltwater from low-latitude (far-field) ocean basins into high-latitude (near-field) regions in order to fill space vacated by the collapse of forebulges at the vicinity of previously ice-covered areas. Total melting of ice reservoirs and thus the resulting eustatic sea-level rise were predicted to end at around 4–5 ky BP (Nunn and Peltier, 2001; Mitrovica and Milne, 2002). The effects of the associated sea-level highstands in the Pacific oceanic islands (e.g. Fiji, Society, Marshall, Caroline, Hawaii) and continental areas (e.g. South China Sea) are recorded by emergent reef structures dated from about 7.0 to 1.5 ka BP (Grossman et al., 1998; Dickinson, 2001; Nunn and Peltier, 2001; Yu et al., 2004a). By contrast, as emphasized by Camoin et al. (1997), in the Western and Central Indian Ocean, recent emer-

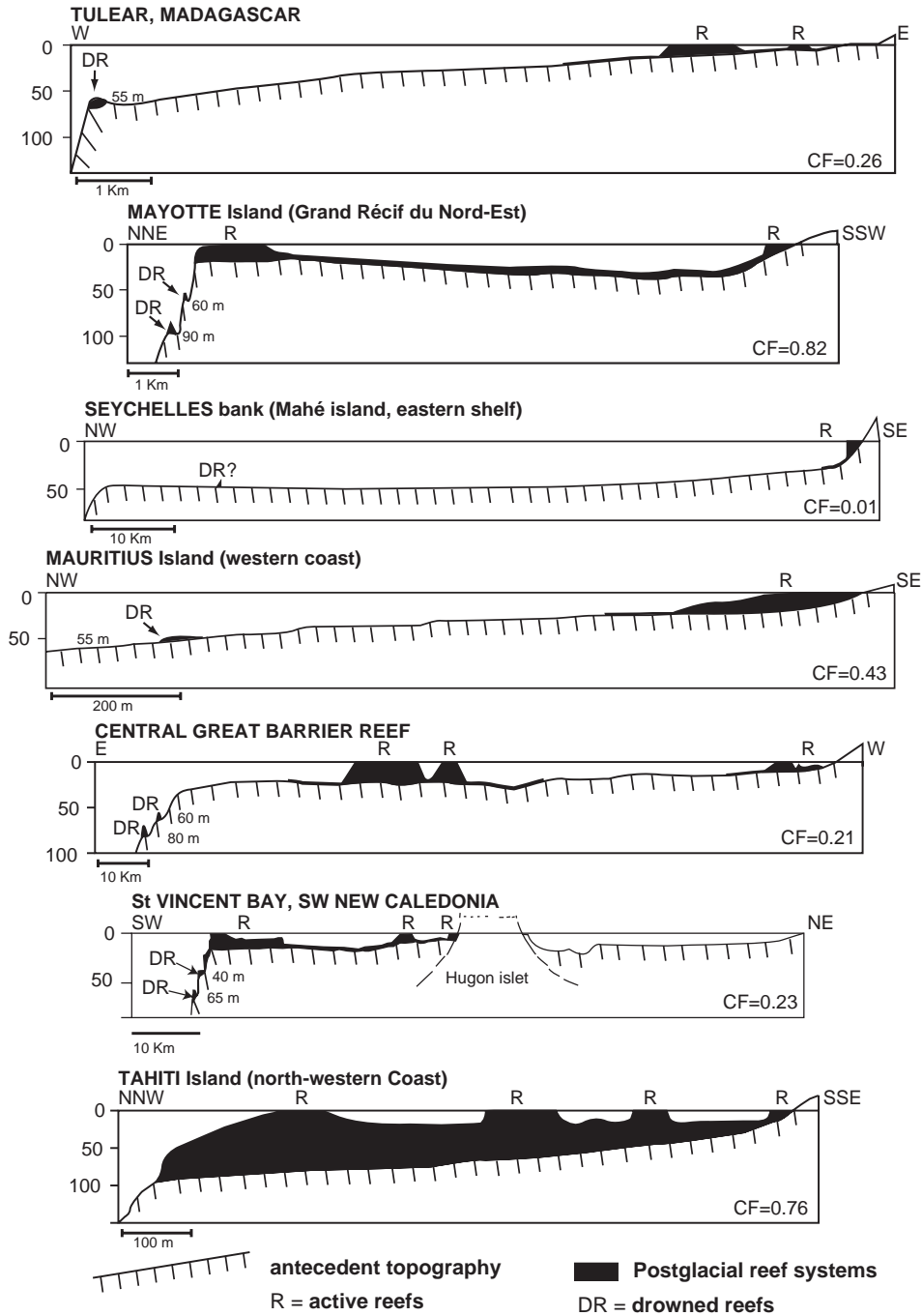


Fig. 12. Topographic profiles of continental and oceanic shelves from the Indo-Pacific province, showing the main morphological features and the areal distribution of the Postglacial reef systems. Coverage factor (CF) refers to the total shelf area covered by the identified reef systems that developed during the deglaciation. It is expressed by the ratio between the area covered by the reef systems of that of their shelf foundations. The more shelf-covering is the reef system, the closer to 1 is the CF. Data from Pichon (1978) (Madagascar); Zinke et al. (2001) (Mayotte); Montaggioni, unpublished (Seychelles); Montaggioni and Faure (1980) (Mauritius); Maxwell (1968), and Larcombe and Carter (2004) (central Great Barrier Reef); Coudray (1976) (New Caledonia), Cabioch et al. (1999a) (Tahiti).



gent in-situ reef buildups are missing from the oceanic islands (e.g. Réunion, Mauritius, Rodrigues, Comoro islands, the Amirantes, the Maldives), suggesting that mid to late Holocene highstands have not occurred here. Such discrepancies between the two oceans may be explained by variations in meltwater redistributions as evidenced by their different sea-level histories since the early Holocene (Camoin et al., 1997). When sea level was at or slightly above its present position, reef islands, including sandy cays on platform reefs and conglomeratic ramparts on barrier reefs and atolls, accumulated. Episodes of reef-island accretion occurred variably from site to site.

Accordingly, in the Pacific, the surficial morphology of many reef flats could partly be the product of erosional processes, mainly biologically driven and acting within the intertidal zone (Trichet, 1969; Guilcher, 1988). Their ages vary from about 6.5 to less than 0.5 ka BP. Most sandy cays (low reef islands) formed in the period from 3 ka BP to present (Tracey and Ladd, 1974; Stoddart and Steers, 1977; Woodroffe, 1992; Woodroffe et al., 1990, 1999; Richmond, 1992; McLean and Woodroffe, 1994).

#### 4.2. Tectonics

Contrasting models of reef evolution are widely assumed to reflect a response to differences in geodynamic setting (Hopley, 1982, pp. 382–384; Scott and Rotondo, 1983; Scoffin and Dixon, 1983). The structural development of recent reefs demonstrates the complex interplay of vertical and horizontal displacements.

On passive margins, long-term subsidence explains the differences in depth to the antecedent foundations across shelves. Davies (1983) reported a uniform subsidence of  $0.05 \text{ mm year}^{-1}$  for the eastern Australian continental margin. Over the 10-ka interval in which the modern reefs have grown, subsidence of the margin has not exceeded 0.5 m and, as a consequence, it has had little effect during the Holocene (Webster, 1999). However, Hopley (1983) concluded that modern outer reefs in the northern and central areas of the Great Barrier Reef required about 1000 years longer to reach sea surface than those on the inner and middle parts of the shelf, due to downwarping of the outer margin. Similar divergent patterns of growth histories can also be shown for the

southwestern section of the New Caledonian barrier reef (Cabioch et al., 1996).

In the southern Great Barrier Reef, unusually deep pre-Holocene surfaces (>30 m) at some locations may be related to active major NE–SW trending faults affecting the continental shelf (Kleypas and Hopley, 1993). Movements along these faults are suspected to have produced uplift of recent reef flats within the last 6 ka.

In the Red Sea, the location and orientation (NNW–SSE) of Miocene to modern fringing, ribbon or atoll reef tracts are clearly an expression of crustal rifting and progressive tilting of horsts and grabens (Braithwaite, 1982a,b; Montaggioni et al., 1986; Purser et al., 1993; Dullo and Montaggioni, 1998). Salt diapirism may also locally provide a control on reef physiography, particularly in the southern Red Sea where reef crests show circular or semi-circular outlines suggesting ascending salt domes (Dullo and Montaggioni, 1998).

For similar reasons, the locations and depths of submerged reef terraces cannot be explained solely as reflecting the positions of stillstands during the Postglacial transgression. For instance, Harris and Davies (1989) noted a lack of correlation from site to site in the depth/height of submerged reef-like terraces within the central Great Barrier Reef and suggested that the settlement of some reefs may have been facilitated by shelf-edge rotational fault scarps.

However, on the time scale of the last deglaciation, gross reef geometry does not systematically reflect differences in regional tectonic history. This is demonstrated by comparing the reef record from the intraplate basaltic island of Tahiti, with those of the forearc andesitic island of Espiritu Santo (Vanuatu), and the Huon Peninsula (Papua-New Guinea) at the tectonically active junction of the Australian and Pacific plates. Major large-scale features are common to all three although they are subject to contrasting vertical motions. Subsidence of Tahiti is at rates of some  $0.50 \text{ mm year}^{-1}$  (Bard et al., 1996), uplift of Espiritu Santo at  $5\text{--}6 \text{ mm year}^{-1}$  (Cabioch et al., 1998), and uplift of the Huon Peninsula is at rates of  $0.4\text{--}5 \text{ mm year}^{-1}$  (Chappell et al., 1993). However, irrespective of the directions and rates of these motions, all three sites exhibit expanded reef sequences; 87 m on Tahiti (Montaggioni et al., 1997a,b), 70 m on Espiritu Santo (Cabioch et al., 1998) and about

70 m on the Huon Peninsula (Chappell and Polach, 1991).

Because the magnitude and rates of eustatic movements between 20 and 7 ka BP were 2 to 50 times higher than those of local tectonic motions, neotectonic effects were likely to have been overwhelmed by eustasy and, as a result, do not seem to have been a major control of reef growth patterns. In the past 7 ka, the impact of tectonics relates only to surface erosion of reef crests and flats in response to emergence.

#### 4.3. Antecedent topography

The idea that the physiography of modern reefs is controlled to a large extent by shelf foundations at various scales has provided the framework for recent antecedent basement theories (Steers and Stoddart, 1977; Hopley, 1982; Purdy and Bertram, 1993; Grigg et al., 2002). The formation of Postglacial submerged reefs at the margins of shelves may have been aided by the existence of favourable topographic features formed by erosional or depositional terraces of earlier stillstands (Hopley et al., 1997). Large-scale topographic features such as elevations of shelf breaks and atoll summits, the general distribution of topographic highs, and the overall inclination of shelves, are regarded as directly constraining vertical net accretion rates. Reefs from the outer margins of continental shelves and mid-ocean atolls would display maximum accretion rates, whereas those on topographically uniform inner shelves would experience minimum rates (Kleypas, 1997). However, data from various reef-bearing shelves in the Indo-Pacific strongly suggest either that the physiography, size and accretion rates of Postglacial reefs can have been little influenced by existing topography or that they can be entirely independent (Hopley, 1982; Walbran, 1994; Montaggioni, 2000). Indeed, most modern reefs in the Indo-Pacific have developed large-scale biozonation of their own making during the last 10 ka. Contrary to Longman (1981)'s assertion, they have already reached a mature stage of development and their anatomy clearly reflects the ability of builders to form framework instead of the underlying topography.

Although many shelves offered large surfaces to transgressive waters, the thicknesses and areal dis-

tributions of the reef tracts supported on them vary widely from site to site. These variations can be quantified by a measure of the shelf area covered by Postglacial reef systems (reef piles plus reef-derived deposits). The coverage factor (CF) is expressed as the ratio between the area occupied by a given reef system and that of its shelf basement. The larger the area of shelf covered by the reef system, the closer CF is to unity (Fig. 12).

There is no covariance between overall shelf relief and patterns of reef accretion. Rates of reef growth are not directly tied to location; inner shelf fringing reefs may have developed at rates comparable to mid and outer shelf reefs as in New Caledonia (Cabioch et al., 1996) and NE Australia (Davies and Hopley, 1983).

Small-scale topographic features such as substrate type, changes in slope and paleochannels were probably more important in facilitating or preventing coral settlement than overall shelf architecture (Webster, 1999; Grossman and Fletcher, 2004). Regional differences in the timing of reef initiation and in growth patterns can be explained in terms of substrate character. Reefs preferentially colonize karst surfaces of limestones and rough lava flows, while unconsolidated sediments and smooth-surfaced metasedimentary outcrops are apparently less suitable (Cabioch et al., 1995). The slope of the substrate may have a direct influence on the composition of pioneering coral assemblages (Webster, 1999). Flat and gently sloping surfaces (<15°) are predominantly covered by fast-growing branching and domal forms, whereas steeper dipping substrates (>40°) attract mainly laminar and slow-growing forms. On steeper slopes, a rising sea level will easily overwhelm the ability of slow-growing corals to keep up. Even if antecedence can be shown to have affected the early stages of reef growth and to locally determine the gross morphology of individual reef bodies, it has not strongly influenced Postglacial patterns of reef accretion.

#### 4.4. Sea surface temperatures (SSTs)

Sea surface temperatures provide an important control on reef growth by influencing the composition and structure of coral communities and regulating the aragonite saturation state of surface waters (Kleypas, 1997). In the Pacific, Grigg (1982) demonstrated that

rates of reef calcification decline as a linear function of increasing latitude, due to the decrease in mean SSTs towards the upper limits of the tropical belt. However, the way in which the tropics have responded climatically to the LGM and deglaciation is still questioned (Fig. 11c). In particular, there is controversy concerning the record of tropical SSTs during the LGM (Mix et al., 2001). In the tropical Indo-Pacific, the LGM/Late Holocene temperature contrast derived from Sr/Ca in corals is 4–6 °C (Beck et al., 1997; Taylor et al., 2000). This conflicts with microfossil-based or modelling reconstructions that give a maximum range of 1.2–3 °C (Barrows et al., 2000; Hostetler and Clark, 2000; Lea et al., 2000; Lee et al., 2001; Steinke et al., 2001; De Deckker et al., 2002; Koutavas et al., 2002; Liu et al., 2002). Revisiting the question of the temperature, size and variability of the Indo-Pacific Warm Pool since the LGM, Gagan et al. (2004) found that SSTs herein were 3 °C cooler than at present during the LGM, while Barrows and Juggins (in press) observed cooling of up to 4 °C in the tropical eastern Indian Ocean. The warm pool is believed to have regulated low-latitude climate throughout the Pacific during the LGM; Thunell et al. (1994), and De Deckker et al. (2002) pointed out that SSTs may have varied by less than 2 °C from those of the present. By contrast, in enclosed epicontinental seas (e.g. the Japan Sea), SSTs during the LGM are considered to have been about 2 °C higher than the present as a result of heavily stratified surface waters (Ishiwatari et al., 2001). Crowley (2000, in Mix et al., 2001) stressed that a 5 °C-cooling would take the temperature below the survival limit of most corals and would have restricted the distribution of reef-building corals during the LGM to only about 5% of its modern range. This contention is, however, open to debate. According to Kleypas (1997), the total area available for reef growth during the LGM was about 20% of the present area, due principally to a decrease in available space (smaller areas on steeper slopes) at the lower sea stand and secondarily to the fall in SSTs. The existence of coral reefs in various Indo-Pacific areas, and particularly in the Indo-Pacific Warm Pool between 23 and 18 ka BP, provides strong evidence for SSTs conducive to reef growth, except maybe in regions such as the eastern equatorial Pacific that seem to have received colder waters from southern

higher latitudes (Loubere, 2001). Reef-building coral communities are found at present along coasts where winter SSTs are less than 18 °C, falling to 13.3 °C and 16 °C in the highest-latitude sites in the Ryukyu Islands (Yamano et al., 2001b) and the northern Persian Gulf (Purser, 1973), respectively. These communities formed Holocene sequences 8 to 20 m thick at rates ranging from 2 to 6 mm year<sup>-1</sup>. As a result, the area occupied by coral reefs during the LGM was probably larger than expected, even if SSTs were depressed by more than 2 °C in comparison with the present and seasonal thermal fluctuations were equivalent to or slightly greater than those seen today.

In the Pacific, the 19 ka BP jump in sea level occurred about 2000 years before the significant decline in marine  $\delta^{18}\text{O}$  (Clark and Mix, 2000; Mix et al., 2001). This suggests a decoupling of the timing or magnitude (or both) of sea-level and isotope records very early during deglaciation, probably caused by a deep ocean cooling (1–3 °C relative to modern bottom water temperature) associated with the jump. If correct, upwelling deep waters may locally have cooled coastal waters sufficiently to turn off reef growth.

From about 20 to 17 ka BP, there was a sharp global warming trend. This warming may have favoured the growth of El-Niño-like events during the considered time interval (Koutavas et al., 2002; Palmer and Pearson, 2003). SSTs rose by 1 to 4 °C above those of the LGM termination (Bard et al., 1997; Kienast et al., 2001; Steinke et al., 2001; Visser et al., 2003). As pointed out by Weaver et al. (2003), some of the rise in SSTs may have originated from reduction in thermohaline circulation and global increase in CO<sub>2</sub>. This warm event seems to coincide with the initiation of RGI reefs. The period from 15 to 10 ka BP is marked by high-frequency SST changes, and more particularly, by the rapid Bølling warming event at 14.8 ka BP (Grootes and Stuiver, 1997; Mix et al., 2001).

Between around 14.3 and 13 ka BP, the MWP-1A discharge and the subsequent jump in sea level were accompanied with a brief sea surface cooling of 1–2 °C compared with the Bølling–Allerød interval (Sonzogni et al., 1998; Gagan et al., 2000). Even if the amplitude of this fall in temperature is correct, it is unlikely that this event was sufficient to have brought about drowning of the RGI reefs.

As the MWP-1A came to an end, there was a rapid warming that lasted until about 12.7 ka BP (Mix et al., 2001; Gagan et al., 2000) and a marked decline in seasonality occurred (Koutavas et al., 2002). SSTs oscillated around 1 °C below present-day values (Sonzogni et al., 1998; Gagan et al., 2000) and this warming event coincides with the settlement of RGII reefs.

From 12.7 to about 11.5 ka BP, SSTs in the Indo-Pacific were depressed by at least 1 °C relative to those in the preceding warming event (Sonzogni et al., 1998; Gagan et al., 2000; Koutavas et al., 2002). The cooling probably coincided with the MWP-1B pulse and most RGII reefs were drowned during this period.

During the Early Holocene, from about 11 to 8.5 ka BP, SSTs increased; in the central part of the Indo-Pacific Warm Pool, the rapid Postglacial rise in SST led the deglaciation to produce near-modern SSTs by about 3000 years (Gagan et al., 2004). This warmer phase coincided with the start-up of the most active reef generation, RGIII. However, during the 7.5–7 ka BP time interval, the winter SSTs were still significantly cooler (2.4–4.5 °C) than at present due to stronger winter monsoons, at least in southeast Asia (Jian et al., 2000). This result there in coral mass mortality about every 50 years (Yu et al., 2004a). The transition to modern SSTs was punctuated by temporary rapid variations and enhanced seasonal gradients. For instance, high-amplitude SST change (2–3 °C) occurred around the Arabian Sea during the summer (upwelling) season on decadal–centennial timescales (Jung et al., 2002). Between 8 and 6 ka BP, the El-Nino (ENSO) pattern became reduced or inactive (Koutavas et al., 2002). The onset of modern ENSO periodicities seems to have occurred about 6 ka BP throughout the tropical Pacific region, with an abrupt increase in ENSO intensity about 3 ka BP (Gagan et al., 2004). While most reefs developed normally, some at the upper latitudinal growth limits suffered a slow-down in coral colonization (Cabioch et al., 1995). The so-called “Holocene climatic optimum” (up to 1 °C above present SSTs) recorded in both the Pacific and Indian Oceans (Sonzogni et al., 1998; Gagan et al., 2000) coincides with the main phase of sea-level stabilization. Locally the thermal structure of the ocean water may have been influenced by the sea-level behaviour. Thus, for the past 7 ka, a number of reefs have been able to reach sea surface, while new

reefs have continued to settle in a variety of areas: New Caledonia (Cabioch et al., 1995), the Ryukyu Islands (Yamano et al., 2001a,b), western Asia (Tudhope and Scoffin, 1994; Yu et al., 2004a), Australia (Johnson and Risk, 1987), and the eastern Pacific (Cortés et al., 1994).

#### 4.5. Nutrient levels

Nutrient supply was probably a major factor in determining the time of reef settlement during the last glacial cycle, and in governing the subsequent development or turn-off of reefs (Hallock and Schlager, 1986; Buddemeier and Hopley, 1988; Marshall, 1988; Montaggioni, 1988; Hopley et al., 1997).

Around a number of ocean basins, mainly as a result of thermocline shallowing and stronger upwelling, nitrate supply to surface waters and productivity locally increased significantly during the LGM (Herguera and Berger, 1994; Farrell et al., 1995; Pedersen and Bertrand, 2000; Muller and Opdyke, 2000). In the western equatorial Pacific, productivity was 1.5 to 2 times higher than at present (Herguera and Berger, 1994) and NO<sub>3</sub><sup>-</sup> concentrations would have been enhanced by 25% to 60%. By contrast, in areas such as the eastern equatorial Pacific (in the vicinity of the Galapagos archipelago) and the Arabian Sea, decoupling of surface and subsurface water chemistry caused a reduction in upwelling, and nutrient levels and biogenic productivity were probably lower than at present (Loubere, 2001; Agnihotri et al., 2003).

Changes in the oceanic nitrate inventory in the tropical Indo-Pacific over the past 23 ka BP are shown in Fig. 11e, and are derived from the <sup>15</sup>N/<sup>14</sup>N ratio (δ<sup>15</sup>N) calculated from bulk bottom sediments. In general, mean δ<sup>15</sup>N nitrate values were relatively low (2.7‰ to 6‰) throughout the LGM, indicating a marked decrease in nitrate reduction at the sediment surface and in the overlying water column. The excess nitrate presumably supported an increased output by intense upward mixing and diffusion of subsurface waters in oligotrophic areas (Pedersen and Bertrand, 2000). During the deglaciation, regional changes in the δ<sup>15</sup>N of organic matter deposited in ocean basins and on adjacent shelf margins can partly be explained in terms of a rapid sea-level rise (Bertrand et al., 2000). The transgression caused a progressive reorganization



of ocean circulation and large-scale upwelling that resulted in nutrient enrichment of shelves and adjacent coastal areas (Marshall, 1988). In areas where the transfer of nutrients from southern high latitudes to the equator was altered during the LGM, the re-establishment of modern thermocline conditions occurred early in the deglaciation (Loubere, 2001). Around 20–18, 13.5–13 and 9–7 ka BP, lower  $\delta^{15}\text{N}$  values in many Indo-Pacific areas seem to have coincided with the onset of RGI, RGII and RGIII. By contrast, the periods of higher  $\delta^{15}\text{N}$  recorded around 14.8–14 and 11.5–11 ka BP in the Arabian Sea (Stuthhof et al., 2001) (Fig. 11e) coincided with MWP-1A and MWP-1B pulses. The intensification of denitrification during jumps in sea level is linked to stronger SW Indian Ocean monsoonal upwelling. This enhanced the flux and degradation of organic matter and resulted in an increasing oxygen deficiency at intermediate water depths (300–500 m). The combination of enhanced upwelling and denitrification may have been responsible for the demise of RGI and RGII. This is consistent with the results of Marcantonio et al. (2001) who used the ratio of  $^{231}\text{Pa}/^{230}\text{Th}$ , a measure of biogenic particle flux, as a proxy for paleoproductivity. Two periods of abrupt intensification of the SW Indian Ocean monsoon were identified, at about 15.3–14.7 and 11.5–10.8 ka BP (Fig. 11e), in which increased wind speeds promoted increasing upwelling. Similarly, upwelling-driven surface biological productivity was enhanced greatly during the early Holocene (Agnihotri et al., 2003). There is indirect evidence of Postglacial upwelling at a number of reef sites. Brachert and Dullo (1991) suggested that during phases of rapidly rising sea level, sharp increases in nutrient levels are reflected in deposition of laminar micritic microbialites along deep forereef slopes. Upwelling and nutrient overload have also been used to explain the remarkable growth of *Halimeda* bioherms at the expense of reef-building corals, at depths of 20–90 m in a variety of Pacific areas (Northern Australia, Indonesia) (Roberts and Macintyre, 1988).

Biotic disruption and the demise of a number of modern Indo-Pacific reefs may have been driven by episodic enhancement of suspended sediment and nutrient loads from terrestrial runoff or groundwater discharge (Montaggioni et al., 1993; Mc Culloch et al., 2000).

#### 4.6. Light levels and turbidity

The depth to which light is able to penetrate the water column varies in relation to latitude and distance from the shore (Kleypas et al., 1999). As light intensity decreases, coral zones become more compressed and the depth at which reefs drown decreases (Hallock and Schlager, 1986). However, the influence of light penetration on the latitudinal distribution of Indo-Pacific reefs remains unclear, as there seems to be no substantial difference in calcification rates in lower- and higher-latitude sites. In high-latitude settings close to mainland areas (e.g. Japan, eastern South Africa), the minimum irradiance required for hermatypic coral growth ( $50\text{--}250\ \mu\text{E m}^{-2}\ \text{s}^{-1}$ ) is restricted seasonally to the upper 7–9-m depth. Many coastal reefs at low latitudes also suffer from low light levels, mainly controlled by high turbidity (Thailand, Indonesia, the Great Barrier Reef, New Caledonia, SW Madagascar, Central America). In the most turbid areas of the Southern Great Barrier Reef, for example, corals only extend to depths of 3–4 m (Van Woesik and Done, 1997). While coral growth can occur at depths as great as 105 m along the outer shelf of the Great Barrier Reef (Hopley, 1994), the lower limit of light saturation compatible with the ability of corals to act as reef erectors in mid-shelf sites is about 20 m (Bosscher and Schlager, 1992). By contrast, on open-sea, mid-Pacific atolls, the minimum irradiance level for coral growth extends to 150–160 m and corals are considered to be able to produce a vertically accreting relief from 35-m depth.

The influence of turbidity on the distribution of corals varies and is related to regional and/or local fluctuations in turbidity regimes that apparently reflect the distribution of muddy sediments (Larcombe et al., 2001). It seems that in the Central Great Barrier Reef, a 5-m-thick coastal muddy wedge deposited during the mid- to late-Holocene is the principal source of sediment in the water column. Wave-driven resuspension of this sediment is the main turbidity-generating process. The muds accumulated in the last 6 ka at mean rates of  $0.5\text{ to }8\ \text{mm year}^{-1}$  and have prograded seawards at up to  $1\ \text{m year}^{-1}$  since sea level stabilized (Larcombe and Woolfe, 1999), rapidly increasing the size of the turbidity reservoir. No living corals are present on the muddy bed, but they develop immediately seaward of the wedge and are not fundamen-

tally different from those in much less turbid zones further offshore. Assemblages are dominated by columnar *G. fascicularis* and branching *Porites* with the maximum cover up to 50%. Such a high coverage in an apparently adverse environment reflects the existence of active currents that periodically transport turbid waters away from the reef or maintain sediment in suspension (Larcombe et al., 2001). However, very short periods (a few hours) of high rates of redeposition occur as calm conditions re-establish after a strong turbidity event. Deposition rates in these areas can reach 3 to 4 orders of magnitude greater than the mean rates of net sedimentation over the last 6 ka. Such drastic changes in turbidity imply that corals growing in highly turbid zones have adopted one of two alternative mechanisms of functioning in response to turbidity stress. Heterotrophy is favoured when turbidity is high, whereas autotrophy is more appropriate under lower turbidity conditions (Anthony and Fabricius, 2000 cited in Larcombe et al., 2001). Kleypas (1996) and Van Woesik and Done (1997) suggested an alternative in the southern Great Barrier Reef where strong tidal flows related to high tidal range may generate turbidity and thus account for the failure of coral communities to establish well-developed reefs. High sediment loadings displaced by tidal flows influence species presence or absence, growth forms and growth rates. Major framework builders, such as branching *Acropora* and domal *Porites*, can be replaced by turbidity-tolerant species such as foliaceous *Turbinaria* and *Pavona* forming single-sided plates. These have the potential for upright growth under low light conditions to form a bafflestone. Kleypas (1996) suggested that the demise event that affected the 8–6 ka BP reefs between 5 and 0.1 ka BP in the southern Great Barrier Reef was caused by an increase in tidal range of at least 1 m. This probably occurred after the outer shelf reefs had reached sea level some 4000–5000 years ago, and resulted in the reduction of mean low water level (i.e. the upper limit of coral growth) and an increase in siltation stress. The Indo-Pacific province includes many coral reefs on turbid shallow shelves where they interact with terrigenous sediments. The reef builders involved usually include massive *Porites*, various faviids and/or foliaceous forms such as *Pavona* (Johnson and Risk, 1987; Tudhope and Scoffin, 1994; Kleypas, 1996; Van Woesik and Done, 1997;

Yamano et al., 2001b; Smithers and Larcombe, 2003). The decline of reefs from Madagascar, Reunion, Mauritius, Indonesia, and the Society Islands, and the delay in settlement seen in Thailand, the Ryukyus, and New Caledonia may also be attributed, at least in part, to turbidity-driven light attenuation and siltation.

Little is known of the extent to which the Postglacial to late Holocene reef decline locally reflects either a reduction in light penetration induced by greater sedimentation. The reasons for the termination of early Holocene reefs also remain problematic. However, in the Indo-Pacific Warm Pool region, the beginning of the monsoonal climate conditions at about 14–13 ka BP was accompanied by a substantial increase in rainfall and a subsequent supply of terrigenous clays by rivers (De Deckker et al., 2002). This may have resulted in a significant increase in turbidity of coastal waters in tropical Australasia. Hopley et al. (1997) assumed that, in the Great Barrier Reef, some shelf-edge reefs were buried by clastic sediments derived upslope. Whatever the conditions were, they resulted in the elimination of fast-growing branching and tabulate framework builders in favour of slow-growing mainly massive colonies, and in a significant decrease in rates of vertical accretion.

#### 4.7. Aragonite saturation and atmospheric CO<sub>2</sub>

The rates of calcification of many autotrophic organisms increase as a function of increasing calcium carbonate saturation and decreasing  $p\text{CO}_2$  (Broecker et al., 1999; Kleypas et al., 2001). Reefal communities flourish where aragonite saturation ( $\Omega$ -arag) ranges between 4.1 and 3.1 (Kleypas et al., 1999). As a result of temperature dependence,  $\Omega$ -arag is usually at a minimum in high latitudes (e.g. Lord Howe island and the Abrolhos islands) and in upwelling areas (e.g. the Galapagos), where CO<sub>2</sub> enriched waters reach the surface. The LGM is assumed to have experienced higher  $\Omega$ -arag values (probably, around 5–6) compared to the present (Buddemeier et al., 1998). During the deglaciation, the aragonite saturation of tropical waters declined significantly to 4.4 (modern, pre-industrial value), in response to an increase in atmospheric  $p\text{CO}_2$  (190 versus 280 ppmv) (Gattuso et al., 1998). Thus, variations in atmospheric  $p\text{CO}_2$  seem to have remained within the tolerance thresholds for reef



calcification. Laboratory experiments suggest that the calcification rates of reef-building corals may not have varied significantly within the range of saturation levels corresponding to the deglacial phase. Moreover, it seems that the phases of increase in atmospheric  $p\text{CO}_2$  may partly have catalyzed reef building, as suggested by the apparent synchronicity between the three main reef growth episodes, RGI, RGII and RGIII, and sharp increases in  $p\text{CO}_2$ . This relation is striking particularly when considering the record of  $p\text{CO}_2$  in the Western Equatorial Pacific (Palmer and Pearson, 2003) (Fig. 11g). RGI reefs started to develop around 17 ka BP, while  $p\text{CO}_2$  increased by about 50–80 ppmv and are coincident with the steepest rise in atmospheric  $\text{CO}_2$  levels (about 330 ppmv) that occurred between 15.6 and 13.8 ka BP. RGII is synchronous with the  $\text{CO}_2$  excursion of 30–40 ppmv between 12.5 and 11.5 ka BP. The highest rate of reefal carbonate deposition (RGIII) occurred after 10 ka BP, in the Indo-Pacific province as  $p\text{CO}_2$  increased abruptly by about 100 ppmv from 8.2 ka BP. Conversely, the episodes of reef growth cessation or of reduced development, respectively, identified between around 14.7 to 13 and 11.5 and 10 ka BP seem to match the sharp declines in  $p\text{CO}_2$  of 50 to 70 ppm that were recorded at about 13.3–12.8 and 10.7–10 ka BP. These observations support the coral reef hypothesis of Opdyke and Walker (1992) that assumes that higher rates of shelf carbonate precipitation favoured increase in atmospheric  $p\text{CO}_2$ .

#### 4.8. Dust input

Shinn et al. (2000) pointed out that the decline in reef vitality, and in particular the widespread mortalities of major reef builders such as *Acropora*, in the last 25 years in the Caribbean, coincides with maximum fluxes of dust from African deserts. Mainly composed of various metals and nutrients, the dust also serves as a substrate for disease-promoting microorganisms. Present-day photoplankton blooms in Hawaii are also regarded as the result of increasing desertification promoting large dust storms in Asia (Chadwick et al., 1999). Similar conditions probably also applied during the LGM that was typified by a drier climate and larger dust input, particularly in the SW Indian Ocean, where dust-fluxes were three to

five times greater than those in the Holocene (Marcantonio et al., 2001) (Fig. 11f). In this region, the cold Younger Dryas interval was also a dustier period. These results are in agreement with assumptions of drier conditions in the Western Pacific during the LGM (Martinez et al., 1997; De Deckker et al., 2002). By contrast, around 10–9 ka BP, there was a slowdown in terrigenous fluxes, reflecting an increasing rainfall in Africa and Asia, from 10 to 6 ka BP (Gasse and Campo, 1994, in Marcantonio et al., 2001). Colder and drier intervals at 24–18.5 and 12.8–11.6 ka BP coincided with dusty events when reef growth was spatially restricted. The onset of RGIII appears to be coeval with a marked decrease in dust supply (Fig. 11f). Even if transPacific and transIndian dust transport during glacial and Postglacial periods was not one of the major factors controlling reef growth, it probably enhanced the effects of inimical parameters such as nutrification.

#### 4.9. Sea-surface salinities (SSSs)

In the Indo-Pacific, sea-surface salinities of coastal tropical seas at present average 34.3‰ to 35.3‰ (Kleypas et al., 1999). Little is known of the salinity range in the last 23 ka. Rostek et al. (1993) pointed out that in the equatorial Indian Ocean, SSSs were 0.5‰ to 1‰ higher than the present, during the LGM and up to about 15 ka BP (Fig. 11d). However, in the eastern equatorial Pacific, as a result of increased precipitation and stronger trade winds, SSSs were probably lower (Loubere, 2001). Salinities declined by up to 2‰ from around 14 ka BP, probably as a result of events leading to the intensification of the SW Indian monsoon at the same time (Marcantonio et al., 2001). Similarly, the Japan Sea would have experienced a substantial drop in SSSs from the LGM to 17.5 ka BP (Oba et al., 1991). Present-day values appear as early as 4 ka BP, as the SW monsoon decreased in intensity. Results obtained in the Western Pacific Warm Pool (Martinez et al., 1997; De Deckker et al., 2002) accord with those of Rostek et al. (1993), at least for the LGM; SSSs were more than 1‰ higher than those today, implying that evaporation minus precipitation was higher over the region. Due to the relative tolerance of reefs to minimum changes in salinity values, and accepting a paleosalinity anomaly of  $\pm 2\%$  relative to present, SSS changes since the

LGM cannot be regarded as a factor limiting reef growth.

#### 4.10. Hydrodynamic energy

Hydrodynamic energy is regarded as a major control on Holocene reef accretion (Bourrouilh and Talandier, 1985; Scoffin, 1993; Blanchon and Jones, 1997; Grigg, 1998; Larcombe and Carter, 2004; Grossman and Fletcher, 2004). In areas subject to destructive open-ocean swell or hurricanes, coral recruitment and subsequent reef building are poor and are restricted below storm-wave base. Reefs are unable to catch-up with sea level. Accretion is generally limited to a veneer no more than 1 m thick or to sparse patches of corals and coralline algae clinging to the antecedent substrate. Growth rates are usually less than  $1 \text{ mm year}^{-1}$ , i.e. below the critical threshold for reefs to accrete normally. Once the reef top reaches the storm-accretion limit, at maximum depths of up to 15 m, the reef can only accrete laterally. At the LGM, hurricanes may have probably been less common and less severe than now in the Indo-Pacific, as a result of colder temperatures. During the deglaciation, with the increase in their frequency and severity, storms may have prevented reefs from keeping-up or catching-up with the rising sea level, favouring reef drowning. This may help to explain the lack of mature fringing and barrier reefs along wave-exposed coasts of many relatively stable, oceanic islands (the Comoros, Réunion, Mauritius, the Marianas, Tahiti, Hawaii, and the Marquesas). The incapacity of various Postglacial reefs, that are currently submerged, to track sea-level rise may also have been due simply to the dominance of destructive forces, bioerosion, sediment transport, and storm-wave damage (Hopley et al., 1997). For instance, in South Africa, the development of modern coral framework is prevented by annually recurring high wave energy events (Riegl, 2001). In addition, hurricanes can stimulate phytoplankton blooms through stirring up deeper, nutrient-rich waters, thus affecting the ecology of tropical, shallow-water benthic communities (Babin et al., 2004). Since stronger hurricanes appear to cause larger blooms, they may have contributed to biologically disturb coral growth during the deglaciation. However, long-term, severe storm activity may locally also have a

positive effect, enhancing rates of accumulation through the replacement of slow-growing by fast-growing corals and “instantaneously” incorporating coarse sediments into the reef pile. In addition, storms by providing episodic nutrient replenishment could partly contribute to coralline reef development (Larcombe and Carter, 2004). High rates of deposition, up to  $37 \text{ mm year}^{-1}$  are reported from a variety of Indo-Pacific fringing reefs, and are interpreted as resulting from rare severe hurricane events (Montaggioni, 1988; Cabioch et al., 1995).

The hurricane-control hypothesis has important implications for reef anatomy and ecology. From a study of bioclastic storm ridge sequences along the Great Barrier Reef, Nott and Hayne (2001) concluded that extreme-intensity hurricanes (central pressures less than 920 hPa) occurred every 200–300 years over at least the past 6.3 ka. In the southern China Sea, Yu et al. (2004b) pointed out that, during the last 1000 years, strong storms operated with an average 160 year-cycle. Such events caused large-scale damage to coral reefs, modifying their structures by placing constraints on the longevity of individual colonies, and changing coral communities. Although reef growth models traditionally emphasize the role of in-place, interlocking, primary framework in the development of wave-resistant structure, drilling results indicate that such a framework is rare in the Postglacial record. Reefs composed of relocated and recemented coral blocks (i.e. secondary framework) or skeletal detritus are more representative worldwide (Blanchon et al., 1997; Hubbard et al., 1998; Braithwaite et al., 2000; Smithers and Larcombe, 2003). Internal reef structure varies as a function of swell strength and/or hurricane frequency (Fig. 13). In hurricane-free regions, primary-framework reefs are able to survive in both high and low-hydrodynamic energy settings. By contrast, in hurricane-swept areas, reef cores are principally the products of reworking and re-deposition of coral debris. In other respects, Larcombe and Carter (2004) proposed to explain the prominent gap in reef growth, that occurred between the flooding of antecedent foundations and the coral settlement at around 10–7 ka BP, in terms of paleocyclonic activity. In many cores throughout the Indo-Pacific, the material atop reef foundations are composed often of coral gravel and rubble, while coral framework facies are found rather at the upper

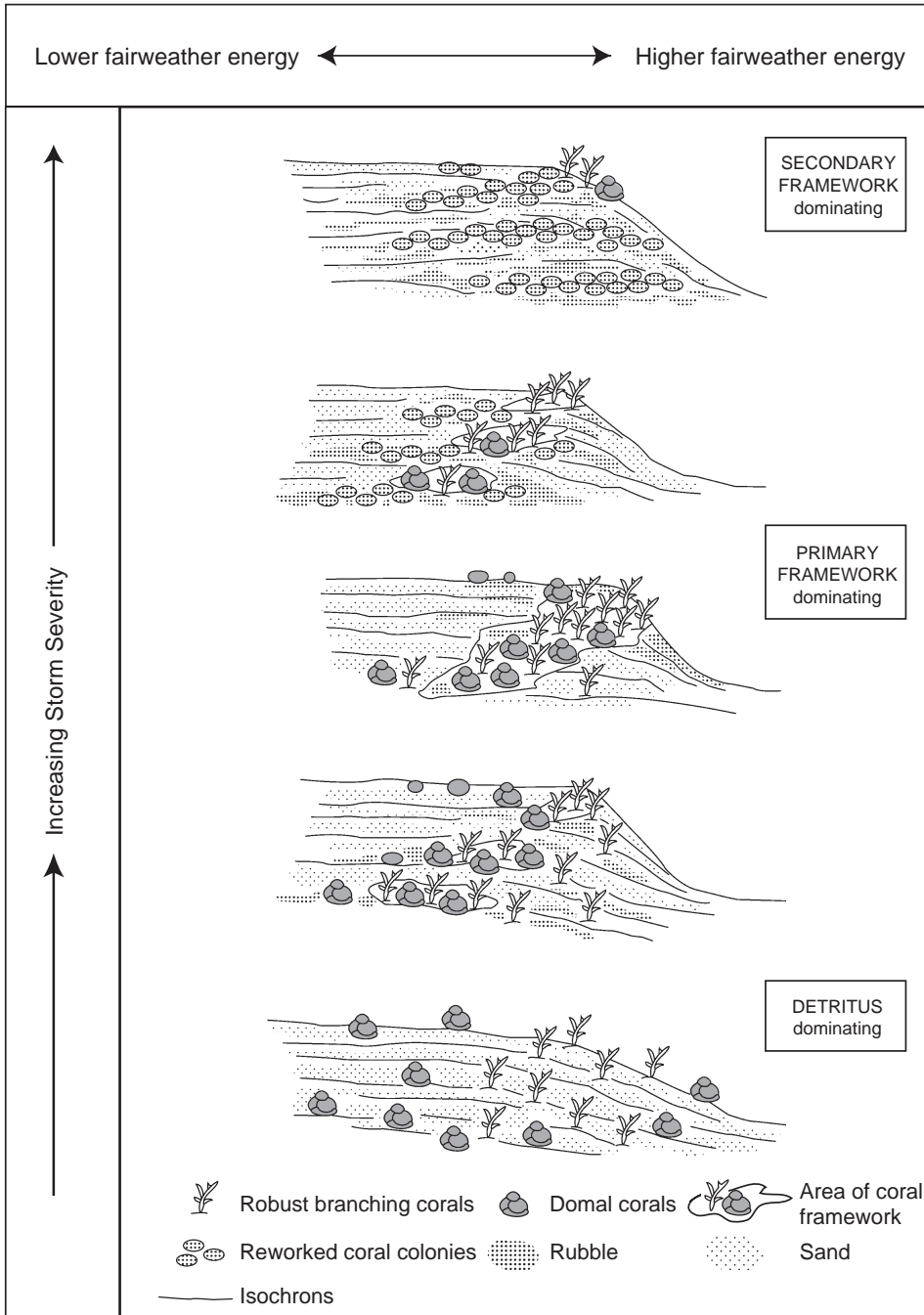


Fig. 13. Generalized models of reef development and internal structure defined on the basis of hydrodynamic energy (modified from Braithwaite et al., 2000).

parts of cored sections. The initial coral growth would have been disturbed by intense reworking in a so-called “cyclone corridor” that would have been traversing shoreward across the location of the modern reef systems as sea level was rising.

The intensity, frequency and locations of hurricanes are expected to respond to climate warming induced by the greenhouse effect (Knutson et al., 1998). Simulations suggest that a sea surface temperature increase of about 2 °C will result in a 5–12% enhancement of hurricane activity. If correct, extrapolation implies that tropical storm activity may have increased by 10–20% since the LGM, making hurricanes more effective in modifying Postglacial reef growth. However, Hayne and Chappell (2001) pointed out that cyclone frequency was statistically constant over the past 5 ka BP in the central Great Barrier Reef while SSTs have increased by 1 °C.

The concept of a high-energy window, introduced by Neumann (1972) and revisited by Hopley (1984), is helpful in explaining some reef growth patterns. It suggests that, in the mid-Holocene, in areas where the rate of sea-level rise outstripped the ability of newly settled coral communities to grow vertically, a phase of higher wave energy may have occurred before a reefal barrier has formed. The window may have remained open from the first complete submergence of the antecedent foundations until the moment at which modern reef flats began to develop, between about 8 and 6 ka BP. The most likely locations of such high-energy environments were those on deeper antecedent platforms (>20 m). Because waves over incipient reefs would have required depths of >5 m to be effective, shallower substrates would have been excluded. Hopley (1984) added that, in reef sites where sea level rose rapidly to its present position (the eastern Indian Ocean and western and central Pacific), at rates of around 10 mm year<sup>-1</sup> between 8 and 6 ka BP, the high-energy window may have had more effect than in areas where the rate of sea-level rise systematically declined from 7 to 1.5 mm year<sup>-1</sup>, between 8 and 3 ka BP. The energy-window effect was surely enhanced during the super-cyclone events reported by Nott and Hayne (2001) and owing to the sediment pumping effect of cyclones, particularly along the cyclone corridors (Larcombe and Carter, 2004). Higher hydrodynamic energy, during fair-weather or cyclonic periods, may have helped in the

rapid winnowing of finer particles from paleosols and terrigenous sediments deposited on inner shelves and coastal zones. It may have limited periods of high turbidity that disturbed reef colonization along many outer shelf margins (the central Great Barrier Reef, New Caledonia, the Seychelles Bank, Madagascar, Tahiti, and the Marquesas). From 6 ka BP, the window progressively closed as reef tops reached a sea level stabilized around its present position and reef flats became sufficiently extended laterally.

Strong tidal currents have also been considered to be a prime determinant of reef architecture. Jones (1995) pointed out that in the Torres Strait (North Australia), reefs developed since about 6 ka BP form 5–10 m thick, elongate detritus-dominated bodies. These developed on previously uncolonized substrate and are aligned to strong regional tidal currents induced by the topographic restriction of the strait.

#### 4.11. *Substrate availability and coral recruitment*

The patterns and timing of coral recruitment after substrate flooding have important implications for Postglacial reef history (Davies et al., 1985; Montaggioni, 1988). Reefs become well developed as a result of high recruitment densities and the longevity of individual colonies, optimizing the persistence of large colonies on site (Chappell et al., 1983). They fail to accrete or to progress beyond the stage of simple coral community or incipient reef, if settlement densities and/or longevities are too low to counteract destructional losses (Van Woelk and Done, 1997).

The attenuation of coral species diversity eastward across the Pacific is primarily controlled by substrate availability (Veron, 1995) but, probably also, by the survival and dispersal capacity of recruits. In the Indo-Pacific province, most of the main reef-building corals are spawners, releasing sperm and eggs simultaneously. The exceptions are the dominant pocilloporid species releasing fertilized larvae (Hughes et al., 1999). Patterns of recruitment differ among the two ecologically distinct groups and species, depending upon stock size, larval survival and settlement behaviour. These disparities may have accounted for differences in rates of substrate recolonization during transgression, because, as stressed by Hughes et al. (1999), reef areas can be sources or sinks for a given coral group. For this reason the inhibition of reef

settlement presumably reflects the absence or limitation of suitable nurseries for particular scleractinian families during lowstands. Shelf edges, banks and seamount summits were the best candidates to serve as refuge sites and centres of dispersal of coral larvae. During the LGM, these potential habitats formed 'islands' separated by thousands of kilometres of unsuitable, non-reef niches. The isolation and spatial limitation of most refuge sites (the Red Sea, Western Indian islands, Maldives, the Ryukyus, and Central Pacific) were incompatible with the maintenance of a high diversity biota.

Recolonization during the transgression required the establishment of patterns of oceanic circulation suitable for larval transport. The large-scale circulation regime in mid-oceanic settings such as high volcanic islands and atolls probably changed little at first, even if the east to west equatorial current regime was less active than to-day. This resulted in the relative impoverishment of coral assemblages in the Central Pacific in particular (Rosen, 1984). By contrast, current patterns near continental shelves were locally completely modified. As stressed by Hopley (1994), the Torres Straits separating Australia from New Guinea closed when sea level was 20 m lower than at present, in the period from the LGM to about 9 ka BP. The closure cut off circulation across northern Australia between the Indian and Pacific Oceans and led to changes in patterns of upwelling. With the establishment of the modern ocean circulation and the re-establishment of exchanges of water masses accompanying, the groups were more easily dispersed.

On scales ranging from local to regional, variability in coral recruitment is driven by large-scale hydrodynamic and meteorological factors, including high water residence times (Sammarco et al., 1991) and latitudinal gradients in water temperature (Hughes et al., 1999). These may differentially affect lateral extension rates of reefs. When integrated over Holocene times, local highly successful coral recruitment may favour carbonate accretion at a higher rate in a given section of a reef.

## 5. Conclusions

This paper is the first attempt to outline a general development scheme for coral reefs since the last

deglaciation on the scale of the Indo-Pacific province. The proposed scheme is based on significant advances in the understanding of the Postglacial evolution of reefs, mainly as a result of borehole studies over the past 25 years and of recent efforts to reconstruct paleoclimate and paleoceanography of the tropics during the last glacial cycle. It appears that reef growth responded differentially to rapid global and local changes in environmental conditions related to the deglaciation. This is expressed through a variety of depositional modes (i.e. aggradation, progradation, retrogradation, downlapping, onlapping) that have generated marked variations in the nature and composition of reef facies, in reef anatomy and in their mutual relationships. The occurrence of alternating or recurrent coral framework facies within most reef-margin piles suggests that rapid changes in bathymetry and water agitation were the principal large-scale forcings in controlling coral population dynamics and reef architecture. However, no single phenomenon can satisfactorily explain changes in reef growth patterns on local to regional scales. Rates of reef accretion were controlled by the synergy of a variety of environmental parameters. The extent to which sea surface temperatures contributed to the inhibition of reef initiation and development is not well known, because it is difficult to separate the effects of temperature changes from those of changes in sea level during meltwater pulses and in nutrient supply linked to upwelling. Contrary to some previous assertions, Late Glacial to Postglacial reef building seems not to have been significantly affected by temperature variability. There is compelling evidence of substantial reef building during the coldest periods of the last glacial cycle. During the LGM, as stressed by Kleypas (1997), reef development was probably more extensive than commonly believed and sea surface temperatures locally have promoted coral reef building. By contrast, nutrient supply seems to have been one of the major controls of reef growth styles, because 'keep-up' and 'catch-up' growth requires constant oligotrophy. Hydrodynamic energy, particularly that reflecting hurricane activity, controlled both growth styles and anatomy. Severe high-frequency storms prevented reefs from tracking sea-level rise, finally helping to promote lateral accretion. Reef anatomy can be related to end-



members of a hydrodynamically controlled spectrum ranging from framework-to-detritus. Substrate availability is the most biogeographically limiting of all physical parameters. The distances between available substrates and centres of coral dispersal, and the alteration of circulation regimes on a regional scale may partly explain variations in the timing of reef settlement and mode of reef growth from one region to another. The degree of influence of changes in atmospheric CO<sub>2</sub> remains unclear, despite the apparent synchronicity between phases of rapid increase in *p*CO<sub>2</sub> and active reef-building events. This is probably due in part to the fact that patterns of calcification have not significantly changed since the LGM.

In summary, nutrient levels, hydrodynamic energy, substrate availability and coral recruitment are the most significant factors controlling coral and reef growth. Other influences, including neotectonics, paleotopography, light levels, dust fluxes, and salinity acted as modulators of these major controls. Thus, as claimed by Hopley (1994), the concept of a tolerance threshold is of prime importance in the analysis of coral reef growth. Beyond a critical limit, a coral community may rapidly evolve towards a coral reef or conversely, may be disrupted, passing from a climatic to a regressive structure.

This synthesis is hoped to provide a starting point for better understanding the processes that set the sensitivity of coral reef systems to environmental changes and for better evaluating reef growth patterns that will be used as monitors to predict future changes. There are critical needs for data on the nature and variability of reef responses to specific environmental parameters. Concerted effort has to be made to obtain high-quality datasets through reef drilling programmes. It is clear that geographic coverage of drilled sites is insufficient for getting meaningful reconstructions of reef growth history on the Indo-Pacific scale. Gaps in sample coverage are particularly acute in the Red Sea, the central Pacific and Indonesia. It is, however, difficult to find and sample thick sequences comparable with those in New Guinea, Tahiti and Vanuatu, particularly when they are submerged. Access to submerged reef piles by coring or submersible diving will provide new insights into the distribution of reef building and growth patterns during the LGM and Late Glacial

times. It may reveal the behaviour of sea level at the LGM-deglaciation and during the mid- to late Holocene, and the impact of increasing storm activity on reef structure in response to increasing levels of atmospheric CO<sub>2</sub>. Similarly, undertaking coring investigations on high-energy outer reef margins and forereef slopes, in spite of logistical difficulty, is required for helping resolve some of the key goals faced in the reconstruction of reef architecture: what have been the respective role for the different depositional processes and growth styles in reef building? For instance, at what extent has back-slope retrogradation contributed to reefs keeping pace with the rising sea level? Are catch-up signatures real facts or simple artefacts of drilling? Achieving these objectives will be aided by process-oriented computer simulations and modelling.

An attempt was made herein to find the best way of distinguishing categories of coral facies based on the identification of the dominant coral builders to specific level. Individual scientists are recommended to evaluate or re-evaluate their core archives in accord with these facies types. The aim is to standardize biological data in order to facilitate intercomparison between sites. In other respects, a full understanding of reef growth history will require further analysis of detrital material (in particular, sand fractions) from cores. Such an analysis was neglected for too long. For example, studies on the changing composition of foraminiferal assemblages through sediment piles will yield additional insights into the environmental conditions driving reef development.

Conversion of radiocarbon ages into calendar dates is recommended to accurately correlate reef development episodes with climatic and oceanographic events and to compare reef growth histories between regions with rigor.

A more detailed geochemical analysis of new and existing cores that sample several centuries of continuous coral growth may provide answers to questions regarding the variability of tropical climates in response to Southern Oscillation events, and changes in nutrient levels in relation to upwelling patterns in the deglacial interval (see Quinn and Tudhope, 2000, for further explanation). However, points of controversy resulting from the use of different geochemical proxies, such as SST in the tropics during the LGM and the deglaciation have to be resolved. In addition,



particular attention to the possible effects of early marine diagenesis on the geochemical signals is also warranted.

Finally, a better understanding of Indo-Pacific reef growth patterns over the last deglacial cycle may aid a more critical application of recent reef models to the ancient record and a retrospective extension of baseline data temporally to approach future change issues more efficiently.

### Acknowledgements

I particularly wish to thank Paul Blanchon, Colin Braithwaite, David Hopley for reviewing an early draft of the manuscript. I benefited from fruitful discussions with my co-workers on Indian and Pacific reefs (Guy Cabioch, Gilbert Camoin, Gérard Faure, Michel Pichon). I also thank Colin Woodroffe for his review of the manuscript. This paper is a contribution to the French National Programme on Coastal Environments (PNEC) financially supported by CNRS, IRD and IFREMER.

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