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Cretaceous history of Pacific Basin guyot reefs: a reappraisal based on geothermal endo-upwelling

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Abstract

The mid-Cretaceous histories (origin, growth, death) of algal–rudist–stromatoporoid reef communities located on many Pacific Basin guyots are complex and controversial. These shallow water, tropical communities originated on volcanic edifices extruded during the Barremian–Albian, grew upward during edifice subsidence/transgression throughout the Aptian, Albian and Cenomanian and several of them died almost synchronously near the Cenomanian–Turonian boundary.

During their periods of origin and growth, we postulate that the reef ecosystems received dissolved oxygen by wave surge and nutrients by geothermal endo-upwelling. By this process oceanic waters of intermediate depth (approx. 500–1500 m) were: (a) drawn into the weathered and fractured volcanic summit and lower part of the older reef and driven upward through the porous framework by the remnant geothermal gradient of the volcanic foundation and (b) emerged at the reef surface to support the high metabolism of the living community.

The death of most of the reefs near the Cenomanian–Turonian boundary approximately coincides with the most intense oceanic anoxic event (OAE) in Pacific Ocean history. During this OAE the chemistry of the endo-upwelled fluids arriving at the reef surface changed from nutrient/oxygen-rich to dysoxic–anoxic–toxic, and killed the community. Additionally, the combination of foundation subsidence and global transgression reduced aeration of interstitial reef waters by wave turbulence.

The post-mortem history of the guyot summits consisted of: (a) drowning to prevent revival of the reef communities, (b) deposition of Turonian–Cenozoic fossiliferous pelagic sediments in the former atoll lagoons and dead reef debris and (c) minor erosion (submarine; karstification?) of the dead reef frameworks.

1. Introduction

During the Cretaceous the Pacific Basin “superocean” (Moberly and Schlanger, 1983; Winterer, 1991, Fig. 2) experienced an unusual, if not unique, congruence of major crustal, oceanic and biologic events that altered its configuration, its ocean and its biota in several fundamental ways. These events included: (a) the production of enormous volumes of basaltic oceanic crust and associated mid-plate edifice-building; (b) an

extended period of rather continuously rising ocean temperatures and sealevels (Arthur et al., 1985; Haq et al., 1987); (c) the development of extensive areas of oceanic dysoxia–anoxia (Schlanger and Jenkyns, 1976; Arthur and Schlanger, 1979; Jenkyns, 1980); and (d) the origin, growth and death of coral–rudistid–stromatoporoid reef communities presently located on the tops of several guyots.

Although the emphasis in this paper is on these communities, their history is strongly linked to

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changes in virtually all of these other aspects of the Pacific Basin. Therefore, an assessment will be made of their relative influences on reef history and a new proposal will be presented to explain the origin, growth and death of many of these reefs.

1.1. Crustal events and Cretaceous reef history: a review

Previous authors (e.g. Menard, 1986; Vogt, 1989; Larson, 1991a, b; Winterer, 1991) have discussed the enormous volumes of basaltic crust that were formed in the Pacific Basin during the Early-mid-Cretaceous (Fig. 1A; column *a*). This crust production was accompanied or succeeded by considerable mid-plate volcanism, (Fig. 1A; columns *b–g*) represented now by numerous guyots in the Mid-Pacific Mountains (Winterer and Metzler, 1984; Van Wassbergen et al., 1990), Geisha Guyots (Vogt and Smoot, 1984) and elsewhere in the basin (Vogt, 1989, Fig. 1). Subsequently, these volcanoes were subaerially weathered and rapidly eroded (1–1.5 km/m.y.; Menard, 1986) by wave-dominated processes. The configuration and relief of the volcanic summit is influenced by the subsidence rate of the foundation, the resistance to erosion of the various rock types (Vogt and Smoot, 1984) and the presence or absence of protective reefs on the volcano margin. The summit relief of the Geisha guyots ranges from 92–256 m ($\bar{X}=183 \pm 58$ m; Vogt and Smoot, 1984) which is in general agreement with the summit relief of such “almost atolls” as Clipperton in the North Pacific, Bora-Bora and Maupiti, French Polynesia and Lord Howe Island, Coral Sea. Installation and upward growth of fringing, barrier and lagoonal patch reefs protects the summit relief until the stage of a pure atoll. The final flat-topped guyot form is only attained after reef atoll drowning when pelagic sedimentation fills the central depressions of the former lagoon (Winterer and Metzler, 1984; Winterer, 1991). It was on and around the margins of these volcanic summits that the earliest reef communities were established during the Barremian (Fig. 1C, columns *w, x*) in the Mid-Pacific Mountains. Some of these reefs clearly reached the atoll stage (Ladd et al., 1974; Winterer, 1991) comparable in most aspects to their modern

central Pacific counterparts. In this area, the reefs flourished from the Early Aptian to the Cenomanian (Fig. 1C, columns *s–w*). The post-reef history of the volcanic foundations included submergence and uplift (columns *d, h*) and final deposition of fossiliferous pelagic sediment (columns *n, o*).

2. Dynamics of modern reefs and oceans

2.1. Geothermal endo-upwelling and reef growth

The geothermal endo-upwelling model of Rougerie and Wauthy has been described, tested and confirmed in several well-documented publications (e.g. Rougerie and Wauthy, 1986, 1993; Rougerie et al., 1991; Andrie et al., 1993). Briefly, the model describes a thermo-convective process in atolls and barrier reefs whereby the remnant geothermal heat from the volcanic foundation drives a continuous upward flow of deep oceanic water through the overlying porous reef framework (Fig. 2). In the modern Pacific in which the majority of oceanic atolls are located, the entering water is Antarctic Intermediate Water (AIW) located at about 500–1500 m depth and characterized by high concentrations of dissolved nutrients (nitrate, phosphate, silicate, CO₂) and elements (Mg, Fe,...). When these waters reach the thin veneer of living algal–coral reef, the nutrients are used to sustain the high levels of primary productivity that characterize these ecosystems. The chemical characteristics of the interstitial waters just below the surficial algal–coral community result from mixing nutrient-rich endo-upwelled water and oligotrophic-aerated oceanic water that is injected into the porous reef framework by wave surge. Symbiotic micro-algae (zooxanthellae) within the coral polyp cells receive these dissolved nutrients by osmosis to maintain their high autotrophic productivity. The reef ecosystem is thus supported by an internal flow of new nutrients, a process similar to an upwelling, but without the well-known consequences for surface water such as phytoplankton bloom, increased turbidity, thermal cooling, etc...

Endo-upwelling maintains the high clarity of the

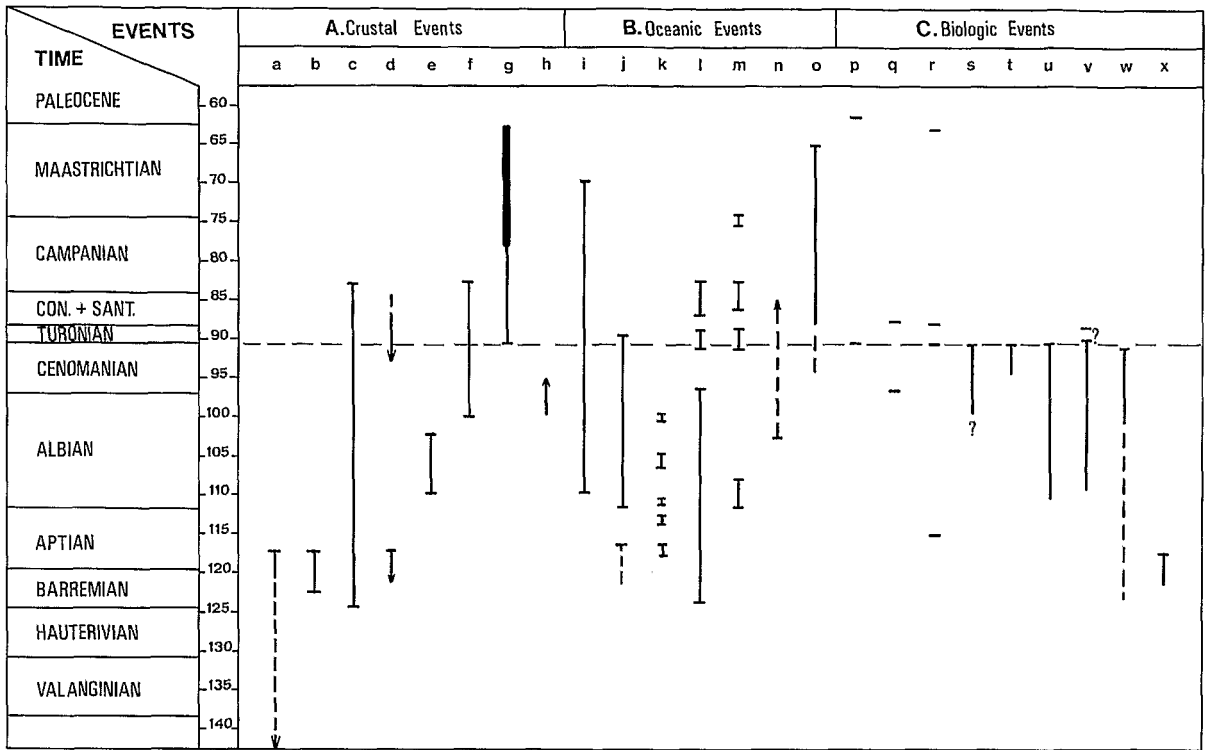


Fig. 1. Summary of mid- to Late Cretaceous crustal, oceanic and biologic events in the Pacific Basin that influenced the origin, location and history of guyot summit reef communities. Note the general congruence of oceanic events (columns *j*, *l*, *m*), mass extinction events (columns *p*, *r*) and the death of reefs (columns *s*–*w*) near the Cenomanian–Turonian boundary (approx. 92–89 Ma) and the absence of congruence at other times, except possibly during the mid-Aptian (columns *a*, *b*, *d*, *j*, *k*, *s*, *x*). Dashed vertical lines represent approximate intervals or extrapolation of data. Arrows on lines, columns *d* and *h* indicate directions of relative sealevel change. a. Origin of plateau basalt crust under Mid-Pacific Mountains (upper solid part; Winterer, 1991) and elsewhere in Pacific Plate (dashed). b. Origin of Mid-Pacific seamounts (Winterer and Metzler, 1984; Winterer, 1991). c. Superplume eruptions (Larson, 1991). d. Subsidence, reef drowning (lower); subsidence (upper) in Mid-Pacific Mountains (Winterer, et al., 1993). e. Origin of Geisha/Japanese seamounts (Vogt and Smoot, 1984; Winterer et al., 1993). f. Origin of seamounts, Wake Group (Van Waasbergen et al., 1991; Winterer, 1991). g. Origin of Line Island seamounts (below); recruitment-death of reefs (above; Winterer, 1991). h. Uplift, erosion, karstification, Mid-Pacific seamounts (Winterer, 1991; Van Waasbergen and Winterer, 1993). i. 35 m sealevel rise, W and central Pacific (Winterer and Metzler, 1984). j. Sealevel rise (Matthews et al., 1974; Larson, 1991, fig. 1; Vogt, 1989, Fig. 3; Winterer, 1991). k. Global OAE's (Bralower and Sliter, 1990; Arthur et al., 1990; see also Hay, 1988, fig. 3). The earliest OAE is the so-called Selli Event. l. Pacific Basin OAE's (Larson, 1991, fig. 1). The OAE beginning at the end of the Cenomanian is the so-called Bonarelli Event. m. Pacific Basin OAE's (Vogt, 1989, fig. 3; Hart and Leary, 1990). n. Pelagic sedimentation on Mid-Pacific seamount summits (Van Waasbergen and Winterer, 1993). o. Pelagic sedimentation on Japanese seamounts (Winterer, 1991; Winterer et al., 1993). p. Mass extinctions, planktonic foraminifera (Hart and Leary, 1990). q. High taxonomic turnover, Caribbean rudistids (Johnson and Kauffman, 1990). r. "True extinctions", Caribbean rudistids (Johnson and Kauffman, 1990). s. Life and death of guyot reefs F, M, P (Matthews et al., 1974). t. Life and death of guyot reefs D, E (Matthews et al., 1974). u. Life and death of Hess, Cape Johnson, Horizon Guyot reefs (Matthews et al., 1974). v. Life and death of Japanese Guyot reefs (Vogt and Smoot, 1984; Winterer et al., 1993). w. Life and death of Darwin Guyot reef (Ladd et al., 1974). x. Life and death Mid-Pacific Mountains guyot reefs (Winterer and Metzler, 1984; Winterer, 1991; Winterer et al., 1993).

ocean, enabling zooxanthellae to receive radiant energy to depths of 50–+80 m. Fissures in the volcanic foundation and reef framework provide conduits into which some of the endo-upwelling

flow may be diverted into sublagoonal sediments to sustain patch reef/pinnacle growth (Guilcher, 1991). Thus, the modern Pacific open ocean reefs receive a continuous input of nutrients by endo-

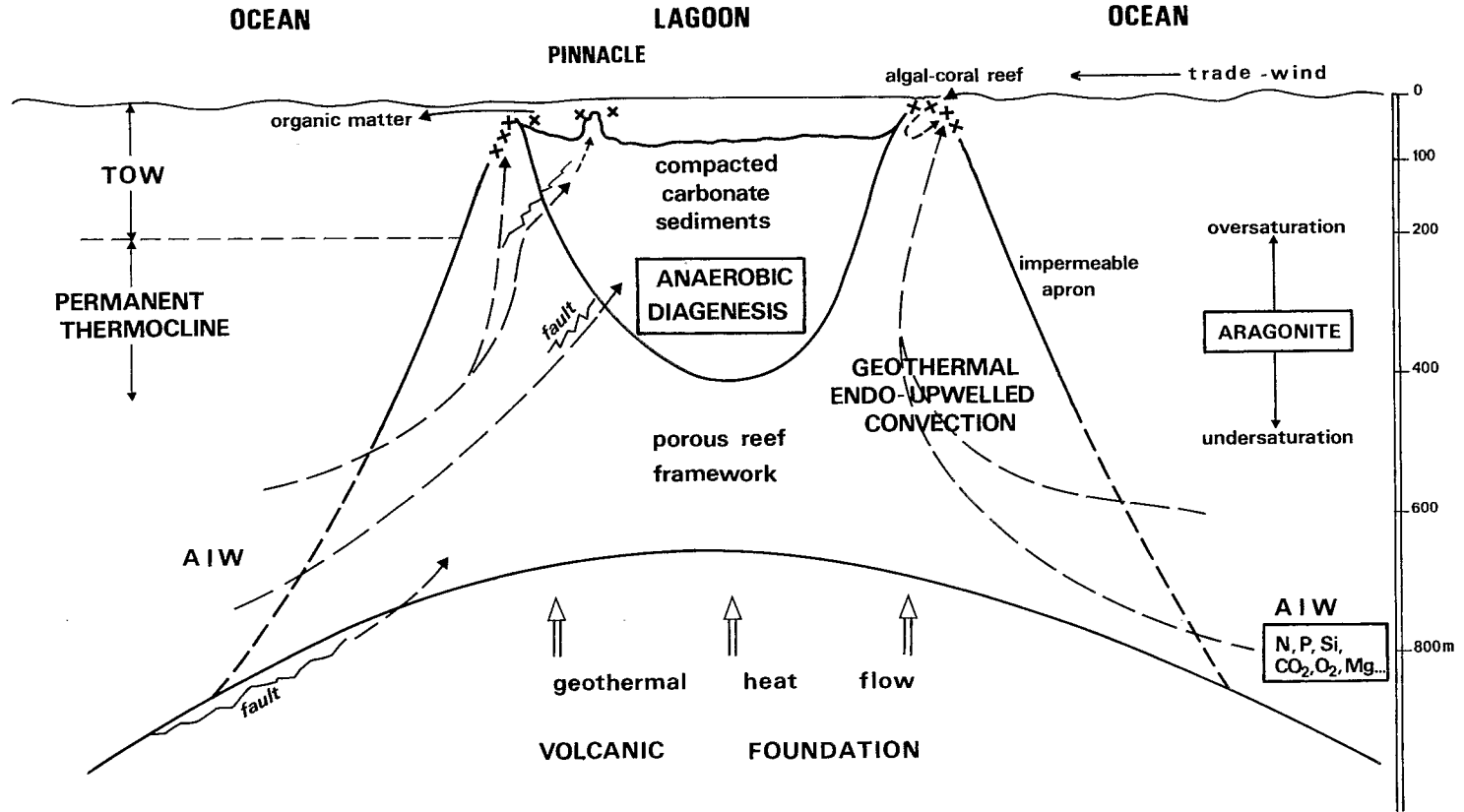


Fig. 2. Generalized cross-section of oceanic atoll showing various aspects of the geothermal endo-upwelling process during an “Icehouse Climatic Supercycle” (Fischer, 1984) as in Fig. 3B. This upward thermo-convection allows nutrient-rich AIW to reach the algal-coral community, sustaining both reef growth (horizontally and vertically as during transgression) and losses (toward lagoon and ocean). *TOW*=Tropical Oligotrophic Water (0–200 m); *AIW*=Antarctic Intermediate Water (500–1500 m); *x*=Location of living algal-coral ecosystem.

upwelling; without this process there would be no reefs able to compensate for the loss of organic matter into the ocean (Rougerie and Wauthy, 1993).

2.2. Modern (icehouse) oceanic circulation and the origin of reefs

Circulation in the modern ocean (Fig. 3B) is mainly governed by dynamic exchanges between deep, cold, nutrient-rich water derived from high latitudes and warm, nutrient-poor shallow water moving toward the poles (Hay, 1988, fig. 5A). Such temperature controlled circulation, called thermohaline, has dominated oceanic circulation since the origin of the present polar ice caps in the Late Eocene–Early Oligocene (Wise, 1987; Hambrey et al., 1989). Growth of these ice caps created a first order perturbation of the pre-existing circulation pattern and of oceanic productivity. Spreading of this cold, deep water (psychrosphere) caused a decoupling within the tropical ocean, the warm surface layer being separated from the deep water, by a steep thermocline. This pattern has persisted to the present oceans in which high latitude downward circulating cold waters are well-oxygenated (about 100% saturated; Levitus, 1983) and in this way aerate deep and bottom layers.

With a well-defined tropical thermocline allowing only low levels of vertical mixing, the warm upper euphotic layer becomes rapidly depleted in nutrients to produce the oligotrophic ocean “deserts” in which are located many of the modern open ocean reefs. For example, commencement of reef growth in the Hawaiian Archipelago was contemporaneous with the origin of the psychrosphere and intensification of the north tropical circulation gyre (Grigg, 1988). It is probable that at the time of origin of these reefs (Oligocene) they were located in oligotrophic waters similar to the modern North Pacific gyre. Their post-Oligocene growth implies a continued decoupling between the shallow euphotic and deep waters, a pattern only provided by maintenance of the polar ice caps. This oceanic decoupling and growth of ice caps is what Fischer (1984) characterized as an “icehouse climatic super cycle”.

The apparently synchronous origin of the Hawaiian reefs coincides with the establishment of

icehouse oceanic circulation and formation of clear, oligotrophic tropical water (reefs cannot grow in nutrient-rich, turbid water or in areas of coastal upwelling for reasons described by Hallock and Schlager, 1986). Once the reefs were established in clear water, they continued to grow, sustained by nutrients provided by endo-upwelling in a manner analogous to the Tahiti barrier reef (Rougerie et al., 1992). Thus, the endo-upwelling model appears to be equally applicable to the origin and maintenance of both barrier and atoll reefs during icehouse supercycles. Can it be tested in living reef ecosystems?

3. Interstitial waters: atoll, barrier and lagoonal reefs

During 1988–1989 five holes were drilled at Tikehau Atoll, French Polynesia (15°S; 150°W) and permanently fitted with polytubes for the purpose of sampling the shallow interstitial water in the reef framework (Rougerie et al., 1990). In 1990 a single hole was drilled in the barrier reef at Tahiti for comparison with the Tikehau waters. In each hole the interstitial waters were sampled at several depths and analyzed for nine basic chemical–physical parameters (Table 1) for testing the endo-upwelling model.

3.1. Pinnacle and atoll

The Tikehau pinnacle is a large patch reef with steeply dipping flanks rising from 20 m depth in the lagoon floor (Fig. 4b). Its flat top reaches the lagoon surface and covers an area of nearly 2000 m², a third of this surface being permanently dry. Due to this outcropping part, a shallow fresh water lens is present during the rainy season to produce a lowering of the interstitial water salinity (Table 1A).

The patch reef waters are mainly characterized by high concentrations of dissolved inorganic nutrients, low pH, very low content (or absence) of free oxygen and negative redox values. By contrast, the interstitial waters of the outer atoll reef contain free oxygen to depths of 30 m and anoxia is present only at 35 m (Table 1B). These

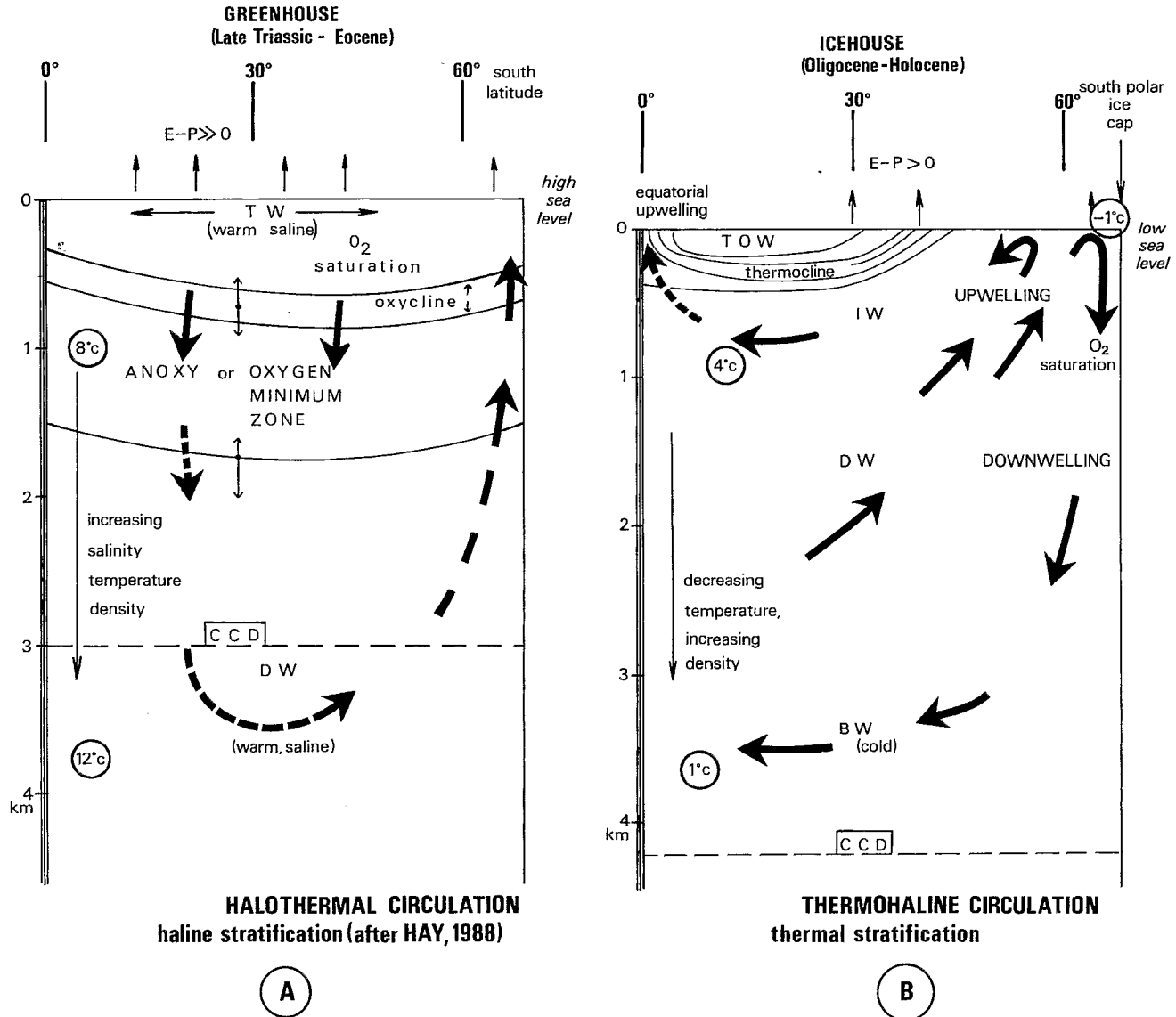


Fig. 3. Comparison of oceanic circulation patterns during greenhouse (A) and icehouse (B) climatic supercycles. *BW*=Bottom Water; *DW*=Deep Water; *IW*=Intermediate Water; *TOW*=Tropical Oligotrophic Water; *E*=evaporation; *P*=precipitation; *CCD*=Carbonate Compensation Depth; *TW*=Tropical Water. The oxycline is the zone of rapid bathymetric decrease in concentration of dissolved oxygen. During greenhouse supercycles sinking of warm, saline *TW* is an ineffective process to aerate deeper water, which rapidly becomes dysoxic-anoxic.

Table 1

Interstitial water database in (A) a lagoon patch/pinnacle reef (1 hole), (B) an atoll reef (4 holes), and (D) a barrier reef (1 hole). Lagoonal (C) and oceanic (E) waters are included for comparison. Mean values (above) and standard deviation (below)

Depth (m)	Salinity (psu)	[NO ₂ +NO ₃] (mM m ⁻³)	NH ₄	PO ₄	SiO ₄	pH	Oxygen alkalinity (1 m ⁻³)	Total potential (eq m ⁻³)	Redox (mV)
A. Tikehau pinnacle (n=40)									
4	35.76	0.37	3.50	0.97	1.90	7.70	<0.5	2.14	-100
	0.14	0.25	2.40	0.21	0.39	0.05	-	0.15	25
10	35.95	0.25	5.20	1.12	2.88	7.57	0	2.05	-180
	0.10	0.09	2.91	0.10	0.80	0.02	-	0.06	30
17	36.01	0.24	6.20	1.25	3.56	7.53	0	2.08	-220
	0.07	0.11	3.12	0.12	1.37	0.09	-	0.11	40
B. Tikehau atoll reef (n=40)									
5–20	35.74	3.05	0.85	0.64	4.17	7.75	1.28	2.04	+112
	0.20	1.60	0.98	0.20	2.40	0.08	0.42	0.16	98
35	35.64	0.88	1.63	1.31	7.10	7.73	0	2.06	+29
	0.08	1.61	1.14	0.27	2.56	0.11	-	0.23	133
C. Tikehau lagoon (n=21)									
0–20	36.19	0.20	0.50	0.22	0.84	8.35	5.0	2.23	+180
	0.19	0.05	0.23	0.11	0.21	0.18	-	0.09	22
D. Tahiti barrier reef (n=20)									
5–20	35.75	1.27	1.30	1.01	22.84	7.70	1.60	-	+130
	0.19	1.05	0.78	0.35	4.45	0.11	0.80	-	85
50	35.70	0.09	10.44	2.31	82.50	7.59	0	-	-121
	0.09	0.05	3.24	0.43	8.42	0.07	-	-	31
E. Ocean (Tropical South Pacific) (n=10)									
0–100	36.20	0.20	0.20	0.20	1.00	8.30	5.0	2.00	+200
500	34.50	25.00	0.10	1.80	12.00	7.90	2.8	2.40	+120

differences in dissolved oxygen result from the surge of oceanic waves which drive well-aerated shallow ocean water into the porous atoll reef framework (Rougerie et al., 1990; fig. 4a herein). Wave turbulence in the lagoon is only important during strong trade winds and cyclonic depressional storms.

3.2. Barrier reef

The distribution of free oxygen in the Tahitian barrier reef interstitial waters is similar to the Tikehau atoll waters, i.e. the upper 20 m are oxic, waters at 20–30 m are transitional and at 30–50 m they are anoxic and contain abundant inorganic nutrients (Table 1D). Thus, the interstitial, shallow (<20 m) atoll and barrier reef waters are a mixture of anoxic endo-upwelled nutrient-rich water and

oligotrophic oxic oceanic water (Table 1E), the result of which is a water with moderate oxygen (0.8–2 l/m³) and significant nutrient content (0.5–1.5 mmole/m³ in phosphate, 2–10 mmole/m³ in nitrate + ammonia).

Conversely, within the Tikehau lagoon, where wave turbulence is much less than in the ocean, anoxia inside the pinnacle extends nearly to the reef surface (Fig. 4b). Consequently corals are much less abundant or absent and sand-gravel covers much of the reef top and leeward pinnacle flanks (Harmelin-Vivien, 1985, pp. 230–232).

3.3. Anoxia and low turbulence

The water data from the Tikehau pinnacle (Table 1A) also indirectly indicate three other

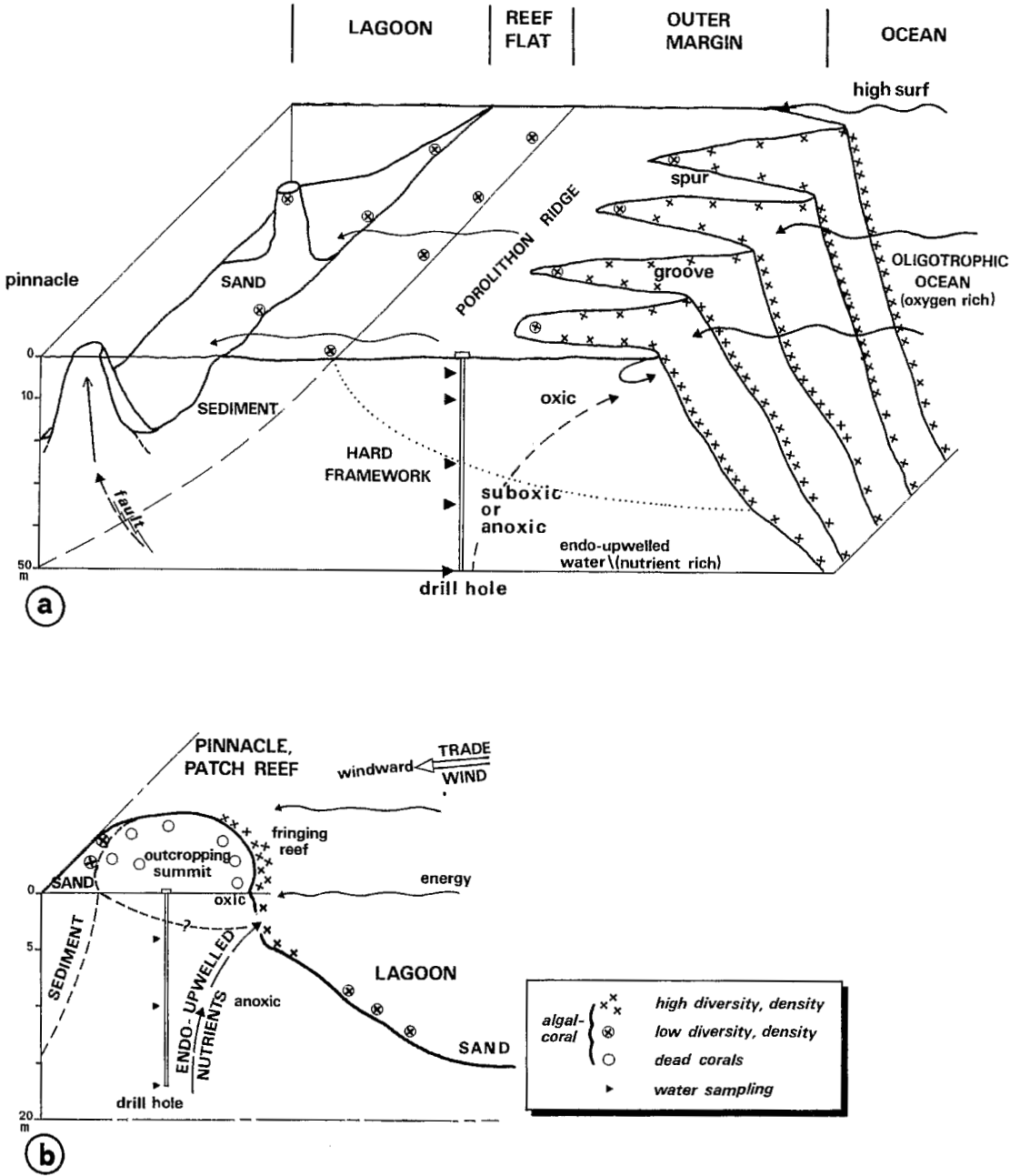


Fig. 4. Schematic diagrams of reef pattern and interstitial water. (a) Barrier reef (high island) or atoll. The outer margin and shallow ocean have a high diversity and density of algae–corals and a well lithified framework. Interstitial oxic layer extends to 20–30 m in drill hole due to strong injection of aerated oceanic water. The reef flat and lagoonal margin have very low diversity and density of algae–corals and is clogged by sediments transported by oceanic waters. Endo-upwelled water enters lagoonal pinnacles via faults (as in Fig. 2). (b) Lagoonal pinnacle or patch-reef. The windward side (NE) has a fringing algal–coral reef community. The leeward side (SW) is clogged by sediment and has an impoverished benthic fauna. Interstitial water is anoxic, except for a thin (4 m) layer on the windward side. In both (a) and (b) the most rapid growth always occurs in the sector where wave energy is highest.

important processes that are the consequence of internal anoxia:

(1) There is sulfate reduction and H_2S diffuses through the reef framework and may locally (as on the leeward side) inhibit larval settlement.

(2) The relatively low pH ($\bar{X}=7.6$) implies that the interstitial waters can corrode the reef framework. This is further supported by the fact that during drilling, the drill stem dropped freely through pores up to 2 m high that were filled with H_2S -rich interstitial water.

(3) The large negative redox potential ($\bar{X}=-180$ mV) indicates that there is active methanogenesis within the framework, as in the Hawaiian patch reef studied by Sansone et al. (1990).

Thus, because of their location in low turbulence environments, lagoonal patch reefs generally are less well-lithified than the barrier and atoll reefs that flourish in high open ocean surf, especially on their windward sides (Guilcher, 1988) where early calcite cementation and vigorous coralline algal growth produce a more dense and durable framework. Early cementation in barrier reefs and atolls is controlled by the CO_2 -carbonate equilibrium; their interstitial waters are of low pH and high total CO_2 but at their interface and mixing with the ocean (high pH; low CO_2) the equilibrium shifts toward carbonate precipitation because of the consumption of CO_2 by surficial algae (zooxanthellae and benthic) and direct degassing. These inter-relations among CO_2 consumption, degassing and carbonate deposition may offer a test for the "principle of maximum cementation" in reef environments (Aissaoui and Purser, 1985, 1986).

The differences in lithification and durability of atoll and barrier reefs compared to lagoonal patch reefs are readily apparent when the effects of typhoons are compared. In 1983 an important typhoon hit Tikehau and caused direct wave damage and dislocation of large corals on the outer atoll margin to depths of 20 m (Harmelin-Vivien and Laboute, 1986). The impact of the typhoon was different on the lagoonal patch reefs and adjacent sedimentary substrates especially on their windward sides where shallow water branching *Acropora* were reduced to rubble. On some windward slopes a few large *Porites* colonies were lifted onto the tops of the pinnacles and on

others they were detached from the steep upper wall or summit and rolled down the slope to the sedimentary substrate. Thus, it appears as if the attachment (by encrustation, early cementation, and coralline algae) of the patch reef corals to their substrate was less secure than for the shallow water (>20 m) atoll margin reefs. Subsequent recovery of the lagoonal reefs should be slower than for marginal reefs due to the clogging effects of sediment on their summits and flanks and destruction by their higher biomass of eroders and borers (Peyrot-Clausade, 1984; Hallock, 1988).

In conclusion, the high concentration of dissolved inorganic nutrients in the patch reef interstitial water promotes algal-coral growth only in areas sufficiently agitated and aerated to clean the substrate and disperse the shallow water anoxia. Conversely, when agitation ceases, the level of the oxia-anoxia contact rises rapidly within the patch reef, as on most of the leeward side of the Tikehau lagoonal reef from which the data of Table 1A were collected.

3.4. Biota

The variation in these physical-chemical parameters is also reflected in variation of such aspects of the benthic biota of Tikehau Atoll as community composition (Faure and Laboute, 1984), substrate cover and species diversity. The coral data in Table 2 were collected less than one year before the 1983 typhoon and therefore are typical of long-term unstressed atolls in the Tuamotu Archipelago. Unfortunately, there are no coral cover data for the lagoonal pinnacle but in our experience such cover is much lower than for the other environments in Table 2. In addition, despite the high coral cover in the spurs and grooves (Table 2B), live substrate cover here generally exceeds that in other environments due to the absolute domination of the surface by encrusting coralline algae, especially *Porolithon onkodes* (Fig. 4a). Conversely, on the reef facing the lagoon, coral recruitment is low, the reef is poorly lithified and tends to be covered by particles carried into the lagoon by ocean water crossing the reef flat (Fig. 4a). The marked reduction in coral species diversity and cover from the atoll margin reefs (Table 2; biotopes 4–7) to environments of comparable depth in the

Table 2
Diversity of hard corals and other important benthic taxa, Tikehau Atoll. Biotopes and coral data from Faure and Laboute (1984); other taxa from Harmelin-Vivien (1985)

Location	Water depth (m)	Coral cover (%)	Coral species diversity	Other taxa (spp.)
A. Seaward slope				
(Biotopes 4,5)	10–25	50–60	15+	Algae, sponges, ascidians (5+)
(Biotope 7)	4–10	40–60	18	Algae (3+)
B. Spurs/grooves				
(Biotope 6)	0–4	5–25	12	Algae (11); sponges
C. Lagoon pinnacle/patch reef				
(Biotope 14)	0–2	(less than	8	Algae (6+)
(Biotope 15)	2–6	biotopes	6	Algae (2+)
(Biotope 16)	6–15	4–7)	6	

lagoon (biotopes 14–16) appears to be partly due to the decrease in wave agitation and accompanying decrease in the interstitial concentration of dissolved oxygen, pH and redox potential (Table 1A). Although the waters in the upper part of the pinnacle contain free oxygen in low concentrations, they are far less than optimal for the success of the immediately overlying coral assemblage. Despite the fact that those corals that survive here all contain abundant symbiotic zooxanthellae and so are autotrophic, their abundance and diversity are reduced, suggesting that other factors (i.e. H_2S , CH_4 , turbid water) are also inhibiting most coral recruitment, survival and organic production. However, these waters may have a positive influence on some of the benthic algae, especially those in the well-lighted upper zone. In our experience, the pinnacle summits and windward flanks are covered by areas of fleshy, filamentous algae, including blue-greens, greens and reds (Harmelin-Vivien, 1985, p. 231) as well as *Tridacna* with symbiotic zooxanthellae.

3.5. Summary

The upper interstitial waters in atoll and barrier reefs contain free dissolved oxygen, resulting from

oceanic wave surge. Interstitial waters of lagoonal patch reefs and large pinnacles are mainly anoxic, except on their windward rims which corresponds to the more flourishing sector of their benthic communities. Oceanic turbulence is thus a major factor for:

(1) Cleaning the algal–coral veneer by removing sediment and maintaining the porous framework of the reef.

(2) Destroying anoxia in the upper layer of interstitial water, eliminating toxicity of reduced molecules and perhaps also promoting early cementation of the framework.

These conditions, in association with the seepage of endo-upwelled flow, account for the fact that atoll and barrier reefs and large pinnacles thrive in high surf regimes. Conversely, at times of very low turbulence, as during rapidly rising sea level, or for reefs located in deeper water (20–80 m), the anoxic interstitial layer may reach the reef surface and, in concert with sedimentation, can cause the algal–coral community to die by suffocation. The larger implications of this scenario are considered below.

4. Mid-Cretaceous Palaeoceanography: Do the models fit?

Evidence from several sources indicates that mid-Cretaceous sea surface temperatures (SST) were unusually warm (Arthur et al., 1985; cf. Weissert and Lini, 1991) perhaps the warmest of the entire Phanerozoic (Arthur and Larson, 1990; Frakes, 1979; Kauffman and Johnson, 1988; Hay, 1988, p. 1938; Barron, 1983) and that the poles may have been ice-free (Barron et al., 1981), preventing the formation of cold deep water. As a consequence of such warm temperatures, abundant crust formation, mid-plate volcanism and active seafloor spreading, mid-Cretaceous sea levels (Fig. 1B, columns *i, j*) were also among the highest of the Phanerozoic (Vail et al., 1977; Haq et al., 1987). The combination of high SST's, possible ice-free poles and high sealevels resulted in oceanic circulation patterns during the mid-Cretaceous that were distinctly different from those described above as thermohaline.

Analysis of numerous cores from the deep sea indicates that during the mid-Cretaceous there were several periods of varying duration when the deep oceanic water contained little (dysoxic) or no (anoxic) free oxygen. These periods have been called oceanic anoxic events or “OAE’s” (Fig. 1B, columns *k–m*) by Jenkyns (1980; see also Fischer and Arthur, 1977, pp. 29–33; Hay, 1988, pp. 1942–1949). The longest OAE was during the Aptian–Albian (subdivided into three subpeaks by Arthur et al., 1990 and into four by Bralower and Sliter, 1990). There were at least two others: at the Cenomanian–Turonian boundary (the most intense in the Pacific Basin; Arthur et al., 1984; Arthur et al., 1988; Vogt, 1989) and near the Coniacian–Santonian boundary (not generally recognized in the Pacific).

4.1. “Greenhouse” oceanic circulation

These and other Mesozoic OAE’s occurred during a “greenhouse super cycle” (Fischer, 1984) when earth climates were warmer than the present and oceanic circulation was *not* driven chiefly by temperature gradients (Fig. 3A). From the viewpoint of palaeoceanography, the warm global climates of the Cretaceous, accompanied by high sea level, deep water anoxia and platform drowning were the consequence of the lack of high latitude downward thermohaline circulation.

During this greenhouse period, high evaporation rates in low- to mid-latitudes resulted in the development of a strong vertical salinity gradient, or halocline (Fig. 3A; cf. Hay, 1988, Fig. 5B), rapid isolation and generally sluggish circulation of the deep water column. With significantly diminished deep aeration, the warm Cretaceous oceans developed an expanded or intensified O₂-minimum, or dysoxic–anoxic layer at intermediate depths (500–1500 m; Fig. 3A; Arthur et al., 1984; Thierstein, 1979; Hay, 1988).

In the Pacific Basin this O₂-minimum zone is marked by black organic carbon-rich sediments on the plateaus and on the summits and flanks of some of the guyots. Contemporaneous sediments on the deep seafloor are pelagic brown-reddish-yellowish zeolitic clays that accumulated in oxic sedimentary environments (Winterer, 1991).

Elsewhere, turbidites indicate that the black sediments were redeposited (eg. Arthur et al., 1984, Figs. 10d, 11; Moberly, et al., 1986), perhaps with their sources on the guyot flanks.

As a result of their isolation and sluggish circulation, the mid-Cretaceous oceans were halothermal systems with warm, saline, oxic shallow water and dysoxic–anoxic intermediate layers (Wilde and Berry, 1982; Arthur et al., 1987; Schlanger and Jenkyns, 1976). Thermal vertical gradients were neutral or even positive, depending on the intensity of the haline gradient. The intermediate waters were also rich in H₂S, NH₃, and CH₄ derived from degradation of organic matter by the metabolism of anaerobic bacteria. In addition, the lower pH and higher Cretaceous temperatures resulted in a generally shallower carbonate compensation depth (Figs. 3A, B; Hay, 1988, fig. 2) thereby reducing the depth below which dissolution of carbonate frameworks may have occurred.

5. Endo-upwelling and Cretaceous reef history

5.1. Sources of error

Because the following discussion centers on detailed timing and sequencing of events shown in Fig. 1, it is important at the outset to acknowledge the nature, sources and magnitude of potential errors in determining them. The quality of the data-base in Fig. 1 is variable and dating uncertainties cannot be easily resolved due to the inaccessibility of the sampling localities on which they are based.

(1) *Geochronologic*. The absolute ages and their relation to Cretaceous stage boundaries are subject to errors ranging from ± 4 Ma for the Albian–Cenomanian boundary to ± 25 Ma for the Barremian–Aptian (Harland et al., 1990; Winterer et al., 1993). The latter is especially significant because the mean length of all Cretaceous stages is only about 6–7 Ma. For dating the Pacific guyots, problems due to thermal rejuvenescence (“secondary volcanism;” Winterer, 1991; Van Waasbergen and Winterer, 1993) may also be involved and influence the crustal dating of events in Fig. 1A, columns *a, c, e* and *g*.

Many guyot events have been dated by palaeomagnetic estimates and comparison with the Pacific Plate apparent polar wander path and have several potential sources of error (Sager, 1992; Winterer et al., 1993). Finally, the value of magnetic anomalies to correlate the events of Fig. 1 is greatly diminished by the fact that they occurred during an extended period of normal polar magnetization (the Magnetic Quiet Chron; 119–85 Ma).

(2) *Biostratigraphic/Chronostratigraphic*. Fossils recovered from guyot summits provide the dates for events in Fig. 1, columns *d*, *h*, *n*, *o* and *s–x*. The precision of these dates is influenced by three factors:

(a) Mixing of fossils by taphonomic processes (reviewed by Scoffin, 1992) and by marine and terrestrial erosional and karstification processes during episodes of uplift and submergence of the guyot summits (Fig. 1A, columns *d*, *h*; reviewed by Van Waasbergen and Winterer, 1993). Such mixing may have involved both reworking of older fossils into younger rocks as well as stratigraphic “leakage” of younger fossils into cracks and pores (Ladd et al., 1974) formed during karstification. Furthermore, Cretaceous reef communities generally lack a well-developed Binder Guild (Fagerstrom, 1987, pp. 422–423) so the dominant frame-building organisms (corals in the Berriasian–Aptian; rudists in the Albian–Maastrichtian) were not securely attached to their substrates. Many of these skeletons were remarkably porous making them easily transported and broken by shallow water processes on guyot summits (Matthews et al., 1974). Dating and sequencing guyot reef history based on samples from mixed locations (e.g. reef rim; central lagoon), fossil ages and communities (reef, level-bottom, pelagic; Fagerstrom, 1964) requires unscrambling to determine which are the most diagnostic for the dating of events or interpreting the palaeoecology of the summit communities, (Ladd et al., 1974).

(b) Many guyot summit fossils were obtained by dredging which may further compound mixing problems.

(c) Fossils that have undergone breakage and alteration by taphonomic, dredging, erosional, diagenetic and karstification processes are usually poorly preserved and difficult to identify

(Matthews et al., 1974; Ladd et al., 1974; cf. Grötsch and Flügel, 1992). Conversely, these processes did not affect to the same degree the distribution and preservation of fossils accumulating on the deep seafloor between the guyots or in the post-reef pelagic lagoonal sediments on their summits and used to date some of the palaeoceanographic events discussed above (pp. 248–249).

These problems are exemplified by the dating of events in the Mid-Pacific Mountains. According to the authors cited in the caption to Fig. 1, the origin of the seamounts (column *b*) and the installation of the earliest reefs (column *x*) were essentially contemporaneous. After the installation and growth of the reefs through the Darwinian subsidence sequence (about 10 Ma; ODP, 1993), they were killed during the mid-Aptian (column *x*) by crustal subsidence and drowning (column *d*). This interpretation of Aptian history contrasts sharply with the Albian–Cenomanian history given by previous authors for reefs D, E, F, Hess, Horizon and Darwin (columns *s–u* and *w*). In the following discussion of guyot reef history, we emphasize the oldest and youngest reported fossils for dating the origin and death of the reefs, respectively.

5.2. Reef origin

The data of Fig. 1A indicate at least four episodes of variable duration for edifice-building volcanism (columns *b*, *e*, *f*, *g*). No reefs are present on guyots in the Wake Group (column *f*) and in the Line Islands (column *g*) reef communities were not present until the Campanian–Maastrichtian (Winterer, 1991). The earliest reefs on the Japanese Seamounts (reefs M, P, column *s*; Cape Johnson, columns *u* and *v*) appear to post-date the earliest reefs in the Mid-Pacific Mountains (column *x*) by several Ma.

5.3. Reef death: cause–effect

The most remarkable feature of Fig. 1 is the near-synchronicity of several oceanic and biological events at or near the Cenomanian–Turonian boundary. They include the onset of the most intense Pacific Basin OAE (columns *l*, *m*), mass-extinctions of foraminifera and rudists (columns

p, *r*) and reef death on several widely distributed guyots (columns *s*, *w*; see also Grigg and Hey, 1992, Table 2). In addition, this was a time of rising sealevel (columns *i*, *j*) and elevated SSTs. We regard this convergence of events as a cause and effect relationship among several inter-related factors, the most important of which for reef history was the Cenomanian–Turonian OAE. This convergence of data also suggests an alternative to the reef death by foundation subsidence and drowning scenario of most previous authors (columns *d*, *x*).

Previous authors (eg. Schlager, 1981; Vogt, 1989; Wilde et al., 1990) have noted the difficulties in linking deep water anoxia, coastal upwelling and crustal events to such biologic events as the death of shallow reef and planktonic communities and to mass extinction or evolutionary radiation. However, by combining elements of greenhouse palaeoceanography (Fig. 3A) and the geothermal endo-upwelling process (Fig. 2), it becomes possible to link the Cenomanian death of guyot reef communities to the Cenomanian–Turonian OAE. Thus, dysoxic–anoxic–toxic water at palaeodepths of 500–1500 m (Thierstein, 1979; Arthur et al., 1984) entered the volcanic foundations and reef flanks (Fig. 5), was driven upward by the remnant geothermal flux and emerged in shallow water (0–50 m) in proximity to the living reef crest and lagoonal patch reef/pinnacle communities. Although the chemistry of these waters shifted to greater anoxia, the thermo-convective process was very similar to the modern Tahiti barrier reef and the Tikehau pinnacle, as discussed above.

By this new scenario, reef death resulted from the combined effects of low concentrations of dissolved oxygen and high concentrations of dissolved toxic substances (H_2S , CH_4) transported upward by endo-upwelling flow as well as produced by bacterial action in the upper part of the reef framework, as in the Tikehau patch reef.

Two additional important factors suggest that these dysoxic–anoxic–toxic interstitial waters reached the living reef surface and were the direct cause of reef death:

(1) The data of Fig. 1 (columns *e*, *s*, *v*) indicate the reefs were recruited in the western Pacific on young hot volcanic pedestals. From this we infer

that the rate of thermo-convection/endo-upwelling circulation was initially very rapid but was diminished at the time of reef death, 10–12 Ma later. Nonetheless, the rate of water circulation in the guyot summit reefs was probably greater than in the modern lagoonal pinnacle at Tikehau, bringing the dysoxic–anoxic–toxic water even closer to the living surface. Still unknown is the rate of change from pre-OAE oxic water entering the reef foundation to anoxic at the onset of an OAE.

(2) The data of Fig. 1 (columns *b*, *d*, *x*) suggest that the guyot summit reefs may have had difficulty catching up with rising sealevel (transgression; columns *i*, *j*). Therefore, with the reef crest in deeper water, injection of oxic water into the framework by wave turbulence must have been even less important than in the modern Tikehau patch reef. This acted to reduce the flushing and aeration of interstitial water by wave turbulence, allowing the dysoxic–anoxic–toxic waters to reach the reef surface. It is important to recognize that any rapid transgression, not compensated by an equivalent “keep-up” of reef growth, is in itself a process tending to limit the aeration of interstitial water and so allow anoxia to invade and kill the reef crest. The beneficial but paradoxical effect of high wave energy on outcropping reefs—coral colonies can be broken by waves but barrier reefs thrive mostly in agitated sectors—is then taken into account in a heuristic way.

5.4. Post-mortem history

Finally, the data of Fig. 1B, C (columns *n*, *o*, *s*–*x*) indicate that reef death was final and succeeded by pelagic sedimentation on the guyot summits. This is also true in the Mid-Pacific Mountains despite the fact their flanks and summits may have passed through the zone of shallow water for potential reef recruitment at least twice (columns *d*, *h*) during their post-mortem history (Winterer, 1991; Van Waasbergen and Winterer, 1993). The absence of revived reef growth during the Late Cenomanian–Santonian may also be attributed to repeated development of O_2 -minimum zones (columns *l*, *m*).

The ages of the fossiliferous pelagic sediments mixed with and overlying the reefs not only indi-

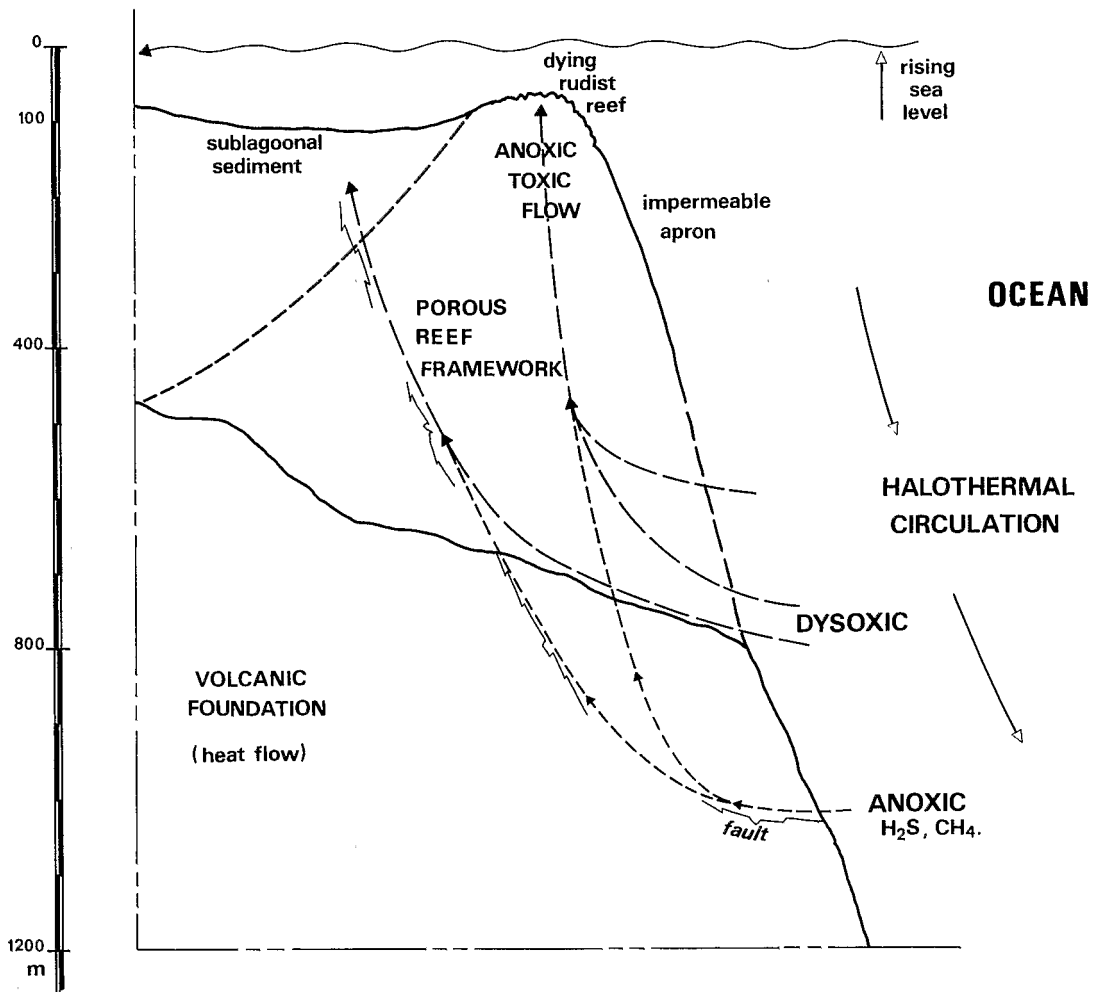


Fig. 5. Schematic diagram of the geothermal endo-upwelling process showing origin, circulation and expulsion of anoxic–dysoxic–toxic waters through a rudistid reef framework (right) and into sublagoonal sediments (left) during a mid-Cretaceous Oceanic Anoxic Event (OAE) as in Fig. 3A. High concentrations of organic matter and sulfur at depth in Enewetok Atoll cores (Schlanger, 1963, p. 1004) support the anoxic–toxic chemistry of the reef framework interstitial waters. Aeration of anoxic interstitial water has ceased as a consequence of the rapid sealevel rise. The rudist reefs are “shot in the back” by anoxic (dysoxic)–toxic endo-upwelled flow.

cate that the interval between reef death and pelagic sedimentation was brief (columns *o*, *v*) but also that the reef communities at the time of death may have been close to the lower bathymetric limit for reef growth (= 80–100 m) based on the possible analogy of coral–zooxanthellae (Stanley, 1981) and rudist–zooxanthellae (Kauffman, 1969; Vogel 1974) symbioses with the various algal–metazoan symbioses in modern reef communities (Cowen, 1983). The absence of revived reef growth also

indicates that during the interval between reef death and the initiation of pelagic sedimentation the combined effects of foundation subsidence and sealevel rise were sufficient to permanently drown the reefs, i.e. they were now below the photic zone. The inability of a drowned reef to catch-up with a rise in sealevel is exemplified by the Marquesas Archipelago, French Polynesia. There is a drowned barrier reef at –95 m around these high islands with no opportunity for coral recruitment. We

believe that this barrier reef system was killed during the Holocene transgression, triggered by the combination of cold water derived from melting of the South Polar and Chilean ice caps and related cooling of the South Equatorial current (Rougerie et al., 1992).

The volcanic and reef history of the Line Island guyots post-dates events in the previous discussion (Fig. 1A, column *g*). Reefs were not installed on their summits until the Campanian and they died near the end of the Maastrichtian (Winterer, 1991) unaffected by Pacific OAE's.

6. Other causes of reef death

Virtually every previous author who has discussed the death of the guyot-capping reefs has also speculated as to the causes of their death (reviewed by Winterer and Metzler, 1984; Winterer et al., 1993). These suggested causes may be arranged under the same headings presented above.

6.1. Crustal events

(1) *Foundation subsidence, uplift; eustasy.* Winterer (1991), Van Waasbergen and Winterer (1993) and Winterer et al. (1993) have carefully documented and described a complex post-Barremian seamount and reef history for the Mid-Pacific Mountains that includes:

(a) Aptian subsidence and Late Aptian reef drowning (Fig. 1A, columns *d*, *x*).

(b) Late Albian uplift exposing the seamount summits and their dead reefs and lagoonal sediments to erosion and karstification and formation of wave-cut terraces on the seamount flanks (Fig. 1A, column *h*).

(c) A second drowning episode beginning in the early to mid-Cenomanian that submerged the summit and flanks (column *d*) and led to pelagic sedimentation beginning in the mid-Turonian or possibly as early as Late Albian (column *n*).

The evidence presented by Van Waasbergen and Winterer (1993) for step (b) is impressive but there are aspects of steps (a) and (c) that leave us doubtful.

If the frame-building corals and rudists were symbiotic with zooxanthellae as noted above, the living mid-Cretaceous reefs were confined to the upper 80–100 m of the water column. In this situation, drowning would occur if the combined effects of the rates of foundation subsidence and sealevel rise exceed the rate of upward reef framework accretion. This was the cause of reef death favored by Ladd et al. (1974), Hamilton (1956), Matthews et al. (1974), Grötsch and Flügel (1992), and others but only Winterer and Metzler (1984) adequately considered the probable rapid rates of upward accretion documented by Schlager (1981) and Fagerstrom (1987) for both modern and ancient reefs.

Winterer and Metzler (1984) estimated that the combined subsidence plus sealevel rise rate during the Aptian–Turonian was about 50–80 m/Ma in contrast to the potential 500–3000 m/Ma accretion rates for modern and ancient shallow water (>20 m) reefs. What is not explained by the previous authors is why the summit reefs after evolving through the shallow water fringing–barrier–atoll sequence as “keep-up” reefs should abruptly “give-up” and drown (Neuman and Macintyre, 1985)? Because the probability of reef death by drowning rapidly increases for progressively deeper water reefs (Hallock and Schlager, 1986), perhaps there were inter-related causes of reef death involving (a) foundation subsidence, (b) reefs living near their lower bathymetric limit, (c) weak guyot-summit currents and slow turnover of ambient water and (d) the arrival of dysoxic–anoxic (the Selli OAE, Fig. 1B, column *k*) endo-upwelled water at the reef crest (the proximal cause of death).

We are also uncertain why guyot summits that formerly supported reefs should not regain their reefs when these summits re-entered the shallow water zone as they were alternately uplifted (step b, above) and flooded (step c, above). The continued endo-upwelling of anoxic water would have prevented such reef recruitment (Fig. 1B, columns *l*, *m*). Finally, the magnitudes of the relative sealevel changes described by Van Waasbergen and Winterer (1993) and Winterer et al. (1993) were considerable in the Mid-Pacific Mountains. If the Late Barremian–Early Aptian reef-building rudists contained zooxanthellae, their mid-Aptian drown-

ing (Fig. 1, columns *d*, *x*) could have required a relative sealevel drop by as much as 100 m. The estimate of the elevation of the uplifted guyot summits (Fig. 1A, column *h*) to produce the erosional and karst features is 180–200 m (Van Waasbergen and Winterer, 1993) which makes the maximum relative sealevel change from a mid-Aptian low to the Late Albian high about 300 m (possibly 1 km; Winterer et al., 1993). Then in the Late Cenomanian–Turonian (Fig. 1A, column *d*) the exposed summits were again submerged below the depth of reef recruitment (–80 to –100 m).

In conclusion, we believe that our verdict of reef death (“by suffocation”) is more efficient and involves fewer problems than “death by drowning”. Our verdict is most firmly supported for the Cenomanian–Turonian events (Fig. 1, columns *l*, *m*, *s–w*) but may also apply to other reefs of the Mid-Pacific Mountains (columns *k*, *x*).

(2) *Drift beyond tropics*. Although the drift rate of the modern Pacific Plate is very rapid (10 cm/yr; Menard, 1986) and may have also been rapid during the mid-Cretaceous, the location of the guyots at 0–32°S latitude (Vogt and Smoot, 1984; Vogt, 1989, fig. 5) and the WNW drift direction of the plate (Winterer and Metzler, 1984) suggest that the reefs were moving closer to the equator at the time of death, rather than away from it. Furthermore, the greater width of the greenhouse tropical belt (Fig. 3A) and the near-synchronicity of reef death mitigate against this suggested cause of death. In fact, Arthur et al., (1984) suggest that the origin of the organic carbon-rich sediments of the Hess Rise resulted from their deposition in the highly productive mid-Cretaceous equatorial zone.

(3) *Burial*. Locally, as at Horizon Guyot, Mid-Pacific Mountains and in the Line Island Group, burial of the summit-capping reefs by contemporaneous volcanic debris may have been the cause of death (Winterer and Metzler, 1984; cf. Van Waasbergen et al., 1990). However, most Pacific Basin guyot summits lack such reef-associated debris and so the hypothesis of death by burial as a general cause of reef death is doubtful.

(4) *Volcanogenic upwelling* (Vogt, 1989). Despite Vogt’s detailed analysis and presentation of this hypothesis, it lacks convincing evidence for the

synchronicity of a major volcanic episode and the death of Pacific Basin reefs with the possible exception of the Mid-Pacific Mountains (Fig. 1, columns *b* and *x*). Additionally, examination of Vogt’s Fig. 3 indicates that the Cenomanian–Turonian reef extinction event occurred during an interval of relatively mild Pacific Basin volcanism (Arthur et al., 1985). Moreover, hot spot theory indicates volcanogenesis and lava spreading to be of local, or regional importance; thus, there is a problem of scale to generalize volcanogenic upwelling to the entire Pacific Basin (Winterer et al., 1993).

6.2. Oceanic events

(1) *Salinity*. Winterer and Metzler (1984, p. 9975) considered the possibility that the salinity of the world ocean was 4–5‰ lower during the Late Aptian and that this caused the death of the reef communities in the Mid-Pacific Mountains (Fig. 1, column *x*). In addition, the generally stenohaline physiology of many modern reef corals lends support to this possibility. Conversely, the sluggish circulation and high evaporation/precipitation ratio in the mid-Cretaceous greenhouse oceans (Fig. 3A) suggests that oceanic shallow water salinities were higher than average. Johnson and Kauffman (1990) and Perkins (1974) suggest that the rudists were euryhaline and thus should have survived moderate salinity fluctuations, either upwards or downwards, during the Cenomanian–Turonian reef death event.

(2) *Oceanic overturn; nutrient excess, etc.* Several previous authors have recognized the near-synchronicity among mid-Cretaceous OEA’s, the death of reefs and/or global mass extinction (Arthur and Schlager, 1979; Arthur et al., 1984; Hallock and Schlager, 1986; Hallock et al., 1988; Wilde and Berry 1982, 1984; Wilde et al., 1990). These authors also recognized the problems of bringing deep water near the surface in order to make the OAE-reef death-mass extinction connection and described oceanographic conditions and causal mechanisms by which such upwelling or overturn would most likely occur. Most of these oceanic conditions and mechanisms (i.e. tectonically-induced marine transgression, nutrient

excess, coastal upwelling) are more applicable to reef death and mass-extinction on continental shelves than in mid-oceans. However, the possibility of mid-ocean overturn bringing anoxic water to shallow water (<100 m) reefs such as those atop Pacific Basin guyots during the Cenomanian–Turonian OAE conflicts with our interpretation.

Arthur et al. (1984) suggested that equatorial upwelling and increased nutrients and organic productivity may have produced the organic carbon concentrations present on the flanks of some Pacific Basin guyots. Equatorial upwelling characterizes icehouse oceans (Fig. 3B) and is trade wind-driven. In mid-Cretaceous greenhouse halothermal-driven circulation (Fig. 3A) there probably was no oceanic equatorial upwelling; all large-scale upwelling was confined to higher latitudes not hospitable to reefs.

Mid-Cretaceous greenhouse oceans belong to Case A oceans of Wilde and Berry (1982) and for the circulation of water masses between the equator and 60° we are in general agreement with their interpretation. In subsequent papers, Wilde and Berry (1984) and Wilde et al. (1990) established the oceanic conditions in which deep water anoxia may develop. Greenhouse oceanic overturn to bring these waters near the surface is most probable during times of important change, especially when entering and leaving glacial climates, i.e. the transition from Case A to Case B (Wilde and Berry, 1982, 1984) or from greenhouse to icehouse climatic supercycles (Fischer, 1984). Although the emphasis of Wilde et al. (1990) is on the possibility of such oceanic overturn causing mass extinction, the implications of mid-ocean overturn during the Cenomanian–Turonian OEA relate directly to the death of Pacific guyot summit reef communities.

Wilde and Berry (1984, pp. 151–152) have noted some special oceanic conditions required for large-scale overturn of anoxic water. In addition, the Cenomanian–Turonian OAE did not occur during the transition from either Case A to Case B or from greenhouse to icehouse (or vice versa) oceans. Although Case A/greenhouse oceans are inherently more unstable than Case B/icehouse oceans (Wilde and Berry, 1984) and thus more prone to overturn, the Cenomanian–Turonian OAE developed during or just after an extended period of rising SST and

sealevel (Savin, 1977; Arthur et al., 1985; Haq et al., 1987) which suggests that stability may have been at a maximum for a greenhouse ocean. In view of these difficulties, we believe that the endo-upwelling of deeper dysoxic–anoxic water is a much more probable cause of Pacific guyot reef death than either massive mid-ocean overturn of anoxic water, upward expansion of the oxygen minimum zone (Arthur et al., 1984) or regional nutrient overloading (Hallock and Schlager, 1986).

(3) *Upwelling/non-upwelling (ENSO) cycle*. As noted above, reefs are never located in regions of coastal upwelling. Similarly, vigorous reef growth is not common in the belt of equatorial upwelling (eg. at Christmas, Canton and Kiribati) because the high plankton load tends to favor heterotrophic filter feeders and borers rather than autotrophic corals. During ENSO (El Niño Southern Oscillation) events, trade winds, westward moving equatorial currents and equatorial upwelling cease and shallow waters become warmer (up to 30°C) and oligotrophic. Such temperatures are close to the upper lethal limit of reef corals which typically respond by expulsion of their zooxanthellae (a “bleaching”) as in the Galapagos during the strong ENSO of 1983 (Colgan, 1990). Thus, on a short term basis the onset of an ENSO event may be comparable in its impact on reefs to the switch from icehouse to greenhouse supercycles. As the data and their interpretation regarding the recent ENSO-non ENSO series become better established, they may become short-term proxies for interpreting longer term climatic changes such as those involving the onset of a Cretaceous OAE or the switch from icehouse to greenhouse oceanic circulation.

6.3. *Biologic events*

(1) *Mass extinction*. Because the Cretaceous was a time of several types of dramatic biotic events, including at least two global mass extinctions with high impact on the marine biota, several authors have sought evidence for various physical–chemical changes in the world oceans as causative factors for these events (e.g. Wilde and Berry, 1982; Wilde et al., 1990; Philip and Airaud-Crumiere, 1991; Johnson and Kauffman, 1990; Hart and Leary,

1990). However, the impacts of these changes on reef communities were mixed in terms of their synchronicity with such presumed biological responses as community composition and diversity (Kauffman and Fagerstrom, 1993), diversity of particular reef-building higher taxa (Johnson and Kauffman, 1990), extinction and radiation events and the sizes and locations of reefs.

In their analysis of the history of Caribbean rudists, including emphasis on the impacts of mass-extinction events on this history, Johnson and Kauffman (1990, p. 318) concluded that the greatest development of rudist-dominated reef communities was during the Aptian–Albian. This proliferation of reefs occurred during one or more OAE's that were best expressed in the Atlantic Basin and their impact on Caribbean reefs appears to have been minimal.

Although the quality of the data base for Pacific Basin reefs and OAE's is much poorer than for the Atlantic and Caribbean, they appear to have had quite a different history and the linkage between reef death, OAE's and mass extinction is closer in the Pacific. Despite the Tethyan similarities of the Pacific guyot summit reef-building rudistids stressed by Hamilton (1956, p. 29), the histories of these communities do not mimick those in the Atlantic for the following reasons:

(a) In the Pacific the Aptian–Albian OAE (Fig. 1B, column *l*) is poorly expressed and appears to have had a possible killing impact only on reefs of the Mid-Pacific Mountains (Fig. 1, columns *k*, *l*, *x*). Elsewhere, many Pacific reefs had their origin and flourished during the Albian (Fig. 1C, columns *s*, *u*, *w*). From this we conclude that during the Albian–Cenomanian the endo-upwelling process continued to bring nutrient-rich oxygenated waters, similar to those of the modern Antarctic Intermediate Water, to most surficial reef communities in the western Pacific.

(b) The record of the Cenomanian–Turonian OAE is quite well-represented in the Pacific, especially in the areas of the guyot-summit reefs (Vogt, 1989; figs. 1, 5). Black shales of this age are indicative of more intense anoxia. They were mostly deposited in moderately shallow water on the ediface flanks; in other cases they were associated with deep water slump deposits (Arthur et al.,

1984). The synchronicity of reef death and the Cenomanian–Turonian OAE stressed above (Fig. 1, columns *l*, *m*, *s–w*), clearly records the change in chemistry of the endo-upwelled water from nutrient and oxygen-rich to toxic and anoxic.

In addition, Pacific guyot reef death coincides with the Cenomanian–Turonian “true” mass-extinction of Caribbean–Tethyan rudists, in which genus and species extinctions far exceed originations (Johnson and Kauffman, 1990, pp. 315–316; Philip and Airaud-Crumiere, 1991, figs. 4, 7). But, unlike Johnson and Kauffman (1990, pp. 317–318) who note the lack of exact synchronicity between rudist mass-extinction in the Caribbean–Tethys Province and most physical–chemical perturbations at or near the Cenomanian–Turonian boundary, we emphasize the strong probability of a cause-effect relationship between the intense Pacific Basin OAE at this time and the death of guyot summit reefs. The probable build-up of deeper oceanic dysoxia–anoxia may not have been synchronous throughout the world ocean, neither was the rate of geothermal endo-upwelling flow nor its transport distance the same in all Pacific guyots. Hence, we would not expect that the Cenomanian–Turonian OAE and reef death to be precisely synchronous on a global scale (see also “Sources of error” above).

7. Conclusions

The data presented above lead us to the following interpretations and conclusions regarding causative factors for mid-Cretaceous Pacific guyot reef death:

(1) Previous authors have not given adequate consideration to the major role of the greenhouse climate and its effect on ocean circulation. The inter-relations among Cretaceous climates, ocean circulation and transgression and the endo-upwelling process are of prime importance in explaining the history of these reefs.

(2) There is near-synchronicity in the time of the death of several western Pacific reefs, the intense Cenomanian–Turonian OAE and associated marine transgression that is neither random nor fortuitous. The “principle of simplicity” (Occam's

Razor) suggests that the endo-upwelling process provides the simplest and hence the most plausible linkage between such deeper water oceanic events as OAE's and shallow water biologic events such as the death of guyot-capping reef communities. The same endo-upwelled flow that had previously sustained reef growth became toxic due to oceanic anoxia (at depth) and lack of aeration (at surface); accordingly, algal–rudist–stromatoporoid communities were “shot in the back” by their internal toxic endo-upwelled waters at the onset of one or more mid-Cretaceous OAE's.

(3) However, interactions among other aspects of ocean history were also involved but to a lesser degree in the history of these communities. These include:

(a) *Sealevel rise; foundation subsidence; water depth.* Following death, none of the reef communities were re-established and pelagic sedimentation dominated guyot summit processes. We believe that this indicates that at the time of their death, the reef communities were in relatively deep water (but still above the aphotic zone). When the stress of OAE dysoxia–anoxia was relieved, the combination of low oceanic wave turbulence to aerate the interstitial water and continuous foundation subsidence brought the dead reefs into the aphotic zone, preventing their revival and re-establishment.

(b) *OAE; mass extinction.* Although the evolution–radiation–extinction of Pacific reef-building rudists is less well-known than in the Caribbean–Tethys, there is an apparent near synchronicity among the Cenomanian–Turonian OAE, reef death and a global rudist mass extinction event. Some previous authors have expressed doubts regarding the importance of this OAE as an agent of mass extinction but we believe that there was a cause-effect relationship among these factors that included the death of guyot summit reefs. Our analysis is largely restricted to the Pacific Basin, whereas the Cenomanian–Turonian mass extinction and OAE were global in scope and affected numerous higher taxa in addition to the rudists.

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Cretaceous history of Pacific Basin guyot reefs: a reappraisal based on geothermal endo-upwelling

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