Geology and Hydrogeology of Carbonate Islands. Developments in Sedimentology 54
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Chapter 15

GEOMORPHOLOGY AND HYDROGEOLOGY OF SELECTED ISLANDS OF FRENCH POLYNESIA: TIKEHAU (ATOLL) AND TAHITI (BARRIER REEF)

FRANCIS ROUGERIE, RENAUD FICHEZ and PASCALE DÉJARDIN

INTRODUCTION

Barrier reefs and atolls of French Polynesia are surrounded by warm and clear oligotrophic water of the South Pacific gyre. Atolls forming the Tuamotu Archipelago show great geomorphologic and hydrologic diversity. The diversity reflects the openness or closure of their lagoons: open, with a pass; enclosed, with or without a shallow pass (hoa); hypersaline on one extreme, or brackish on another; filled with sediment; tilted and/or uplifted. Reef-lagoon systems of other archipelagoes (Society Islands, Austral Islands) are similarly diverse. In all these cases, however, barrier reefs maintain a set of uniform morphological features and must be viewed as the first-order structure. Lagoons and pinnacles range from being second-order structures to being nonexistent.

Interstitial water samples obtained from drillings made in Tikehau Atoll (Tuamotu) and Tahiti barrier reef allow us to describe and produce a model of the behavior of the interstitial water in the reef. This model, named geothermal endo-upwelling, is based on an upward circulation of interstitial water (provided by deep oceanic water), from the reef foundation to the reef rim, and is supported by measurements of nutrients and conservative markers. In the top part of reef and atoll structures, fresh and brackish groundwater are included in the interstitial circulation and govern the dissolution/precipitation balance of the carbonate matrix. Diagenetic processes (early cementation, dolomitization, phosphatogenesis, and degradation of organic matter) are affected by this upwelling of interstitial water.

REGIONAL SETTING

Geography

French Polynesia is mainly a maritime domain that extends over 5 million km^2 in the Central South Tropical Pacific (Fig. 15-1) Encompassed within this area is the Exclusive Economic Zone, which extends 200 miles outward from the islands shores, and is the domain over which the coastal state exercises sovereign rights for exploration, conservation and exploitation of resources (cf., Atlas of French Polynesia, 1993). French Polynesia consists of five archipelagoes, totaling 122 islands and 3,521 km^2. Population is 200,000 with an annual increase of 2.7%.

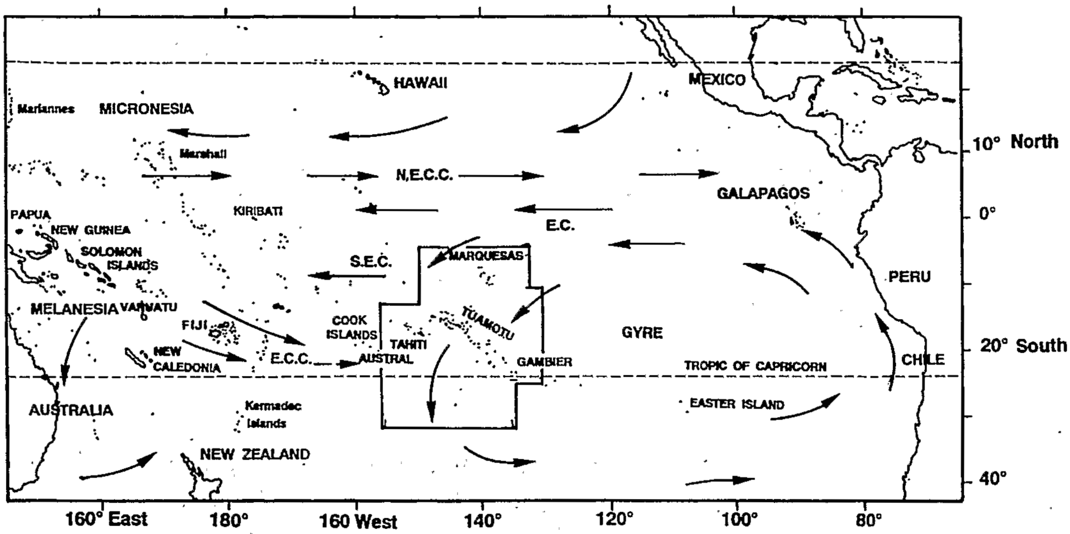

Fig. 15-1. Map of the Intertropical Pacific. French Polynesia is bathed by the oligotrophic oceanic gyre centered in the Easter island zone. Currents flow westward and counter currents flow eastward. Abbreviations: E.C., Equatorial Current; S.E.C., South Equatorial Current; N.E.C.C., North Equatorial Counter Current; E.C.C., Equatorial Counter Current.

Tahiti, the main island, is in the Society Islands archipelago. Its surface area is 1,042 km^2, and it is surrounded almost totally by a barrier reef, separated from the shoreline by a narrow lagoon (100–800 m wide). The Tuamotu Archipelago is the territory's most widely scattered group, a cluster of 77 coral islands (atolls). The most extensive island group of its kind in the tropical Pacific, the Tuamotus stretch over a distance of 1,800 km from northwest to southeast and cover an area close to 1 million km^2 of ocean. The total area of emergent land (or motu) is less than 1,000 km^2, and atolls and lagoons together barely amount to 20,000 km^2. The smallest atolls (with their lagoons) do not exceed 2 km^2, and the largest (Rangiroa) is 1,600 km^2.

Geology

In the Society Islands archipelago, islands show every stage of transition from the original hotspot of volcanism (Mehetia), to the high island with barrier reef (Tahiti), the almost-atoll (Bora-Bora), and the atoll (Tupuai). These islands are moving northwestwards by the movement of the Pacific Plate, which explains why the island groups lie parallel to each other. The linear island chains are formed by isolated volcanic structures set on a plateau lying above the ocean floor. The age of islands increases with distance from a hotspot. The height of emerged volcanoes (2,200 m in Tahiti at present) diminishes with distance from their originating hotspot. The volcanoes finally disappear at the end of the chain, subsiding into atolls and then seamounts. Between birth and atoll stage, approximately 5 m.y. elapse, but atoll growth can keep up with subsidence for tens of millions of years (50 m.y. for atolls of the western Tuamotu).

The chemical composition of the lavas is either tholeiitic or alkaline basaltic. The volcanic flows first discharge in an underwater environment and then in an aerial one. Aerial lava flows commonly contain red layers which are interpreted to be paleosols. The rocks are frequently leached by meteoric water and contain intrusions such as dikes, sills, domes. They have moderate hydraulic conductivity of 10^{-8} to 10^{-6} m s^{-1} (Guille et al., 1993). The sediment cover comprises a transition zone containing volcaniclastic rocks, derived from the weathering of the volcano, and carbonate rocks, derived from the deposition of a chlorozoan assemblage (algae and corals) during the island subsidence and/or sea-level rise. The barrier reefs, which build up along a vertical plane from the rim of the original volcano, contain essentially porous and permeable limestones (hydraulic conductivity of 10^{-4} m s^{-1}), and, in some instances, these limestones have been dolomitized. After total subsidence of any basaltic peak below sea level, atoll morphology represents an unstable balance between chlorozoan construction processes and destruction processes such as mechanical and chemical erosion, slope slides during typhoons, and tsunami. At some localities, destructional processes exceed constructional ones, and the atoll is said to "fail"; it becomes a drowned atoll or guyot. At other localities such as Makatea [q.v., Chap. 14], lithospheric swelling related to the emergence of Tahiti and Moorea 150 km in the south has resulted in tectonic uplift. A bibliography of geology and geophysics of French Polynesia and of other islands of the intertropical Pacific is presented elsewhere (Jouannic and Thompson, 1983).

Climate

Polynesia is under the influence of the Southern Oscillation, a climatic process which involves the interaction between a high-pressure system (centered on the Tahiti Island-Easter Island area) and a low-pressure system (centered on the equatorial north Australian-Indonesian area). Pressure imbalance between these systems produces the trade winds. These winds, averaging 10–20 kn, blow mainly from the northeast sector in austral summer and from the southeast sector in winter. As the high-pressure system is broadly stretched along the subtropical Pacific, between Kermadec Island and Easter Island, it controls both types of trade winds. The northeastern and the southeastern trade winds converge, creating a zone of doldrums and high precipitation, the South Pacific Convergence Zone (SPCZ). Seasonal shifts in the location of the SPCZ are the main cause of the occurrence of a rainy season in the tropical South Pacific. In French Polynesia (Figs. 15-1, 15-5), the rainy season occurs during the austral summer and affects an area between the Tuamotu and Austral archipelagoes. In the Tahitian province (17°30′S, 150°W) mean rainfall at sea level is 150 cm y^{-1}. In central and eastern Tuamotu, rainfall is below 100 cm y^{-1}, and measured evaporation is in the range 150–250 cm y^{-1}. Precipitation minus evaporation (P-E) is hence around −50 cm y^{-1}, a value indicative of marked aridity. In the vicinity of the high islands of the Society and Austral groups, P-E is negative during the austral winter. In the first quarter (austral summer), abundant rainfall and orographic effects on the slopes of the islands lead to positive P-E values. In the Polynesian province, air temperature is generally in the 20–33°C range.

Oceanography

In its eastern and central part, the South Pacific Ocean can be simply described as a large gyre centered on Easter Island and longitudinally limited by the equatorial zone (north of Marquesas Islands) in the north, and the Tropic of Capricorn. Circulation of surface water masses within the gyre is anticyclonic (anti-clockwise) and directly sustained by the trade wind stress (Levitus, 1982). In the surface layer, the dominant current is the Equatorial Current which is oriented westward and flows with an average speed of 20–50 cm s^{-1} (0.5–1 kn) between the latitude of 10°S and 4°N. Another current, located north of the Tropic of Capricorn and known as the South Equatorial Current, flows in the same direction with an average speed of 10 cm s^{-1}. Between these two geostrophic currents, counter currents flow eastward with generally slower, non-permanent fluxes. The Equatorial Counter Current, for example, runs along the SPCZ and brings warm, low-salinity waters originating from the Solomon Sea toward French Polynesia. This counter current is well developed in austral summer and tends to migrate during the winter towards 10°S. The Tahitian zone, located in the western branch of the tropical gyre, has sea-surface temperatures of 26–30°C and sea-surface salinities of 35.6–36.3 psu (practical salinity unit). The high value is a maximum for the entire Pacific Ocean and is typical of the Tuamotu-Pitcairn area where negative P-E differentials tend to increase surface salinity (Delcroix and Henin, 1991). Conversely, in counter currents, surface salinity may decrease to 35.5 psu during the peak of the summer rainy season. Along the Tropic of Capricorn, seawater temperature is around 21°C in winter, largely above the 18°C lower lethal limit for tropical hermatypic corals.

Vertical thermal profiles (Fig. 15-2) show the top 0–150 m of the ocean, labeled "mixed layer," as being directly influenced by air-ocean exchanges, occupied by a warm (>25°C) and saline (36 psu) water mass commonly called the Tropical Surface Water (TSW). This oceanic water is directly in contact with the barrier reefs and atolls of French Polynesia. The TSW has some specific chemical properties such as very low nutrient concentrations (inorganic dissolved phosphate and nitrate below 0.2 mM m^{-3}), saturation in dissolved oxygen (4.5–5. L m^{-3}), high pH (8.3), low chlorophyll concentration (<0.1 mg m^{-3}), low gross primary production (<30 gC m^{-2} y^{-1}) and very low turbidity resulting in the extension of the euphotic zone to a depth of 150 m (Wauthy, 1986). The very low level of net primary productivity is typical of oligotrophic ocean waters, and the South Pacific gyre is often compared to a desert (Barber, 1992).

Below the oligotrophic mixed layer, the permanent thermocline marks a significant drop in temperature between depths of 150 and 400 m. The drop is a transition from warm (>25°C) TSW to cold (5 ± 2°C) Antarctic Intermediate Water (AIW), the latter extending between depths of 500 and 1,500 m. This transition has a major hydrologic

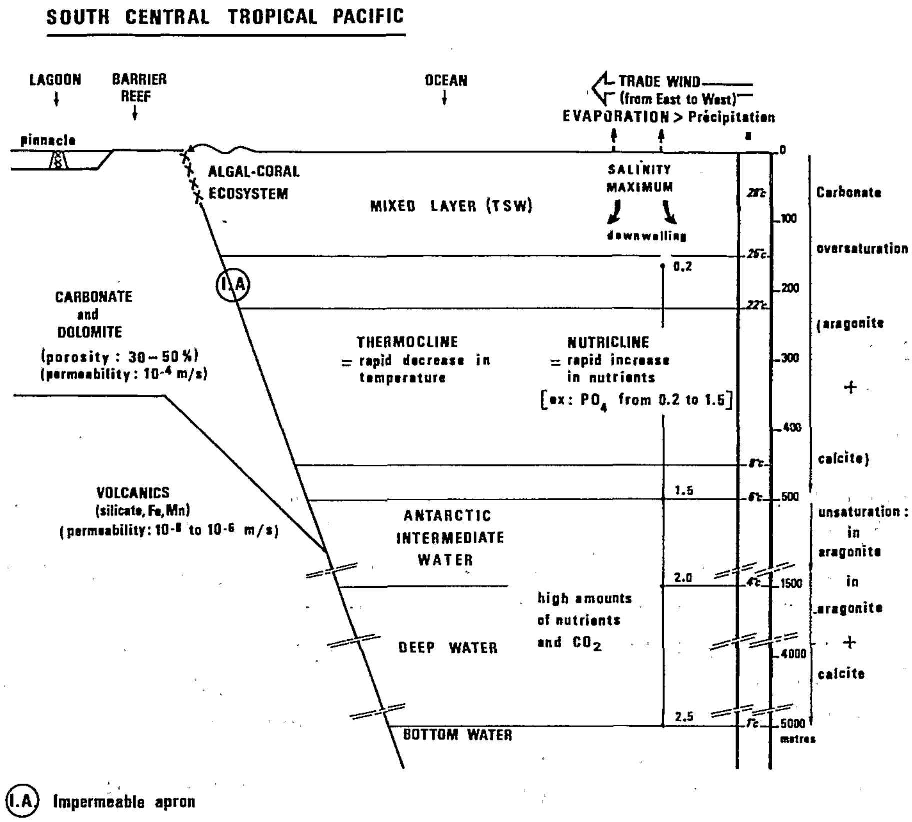

Fig. 15-2. Oceanographic, climatic and geologic setting of atolls and barrier reefs in French Polynesia. Oligotrophy of the mixed layer (0–150 m), occupied by Tropical Surface Water (TSW), is maintained by downwelling of saline surface water, thermal stratification at 150–500 m, and the great depth (> 200 m) of nutricline. High productivity of algal-coral ecosystem in such clear, low-productivity seawaters constitutes the "Darwin paradox" for which we propose a new solution – the geothermal endo-upwelling concept (Rougerie et al., 1992a). The range of the impermeable apron corresponds to the oceanic layer 0–500 m, oversaturated with respect to both calcite and aragonite.

impact because AIW is rich in nutrients (2 mM m^{-3} in phosphate, 20 mM m^{-3} in nitrate). The first kilometer of the tropical Pacific Ocean can thus be viewed as a two-layer system, separated by a permanent thermoclinic barrier: the warm and nutrient-depleted TSW (mixed layer) overlying the cold and nutrient-rich AIW. Oligotrophy is a consequence of that permanent water stratification, and there is no local or regional upwelling to push nutrient-rich water toward the surface, even in the vicinity of the islands (Rancher and Rougerie, 1993; Rougerie and Rancher, 1994). The weakness of turbulences and the thickness of the warm mixed layer prevent any dynamic turnover between the oligotrophic euphotic zone and nutrient-rich intermediate waters (Heywood et al., 1990). The tide range is only about ±15 cm in the Tahiti-central Tuamotu zone.

Upwelling zones in the Pacific basin are located along the American coast (Peru, California) and along the equatorial band from the Galapagos (permanent upwelling) to New Guinea (non-permanent upwelling). The surface-water signature of any upwelling is well known: cool sea-surface temperature anomalies, high nutrient and chlorophyll contents, and enhanced turbidity. It is interesting to note that such properties, which are highly favorable to planktonic development and fisheries, are not favorable to coral settlement and growth (Hallock, 1988); this is the reason why barrier reefs are absent in the coastal upwelled waters from Peru to Mexico and around the Galapagos Islands. Conversely, the oligotrophy of the South Pacific gyre is enhanced by a downwelling process (sinking of surface, highly saline water) with the apparent paradoxical result that atolls and barrier reefs thrive best in clear, nutrient-depleted waters.

GEOMORPHOLOGY

Barrier and atoll reefs

Polynesian patch, pinnacle, barrier or atoll reefs share some general patterns with other Indo-Pacific reefs (cf., Proc. Fifth Intern. Coral Reef Congress [Tahiti], 1985). The reef crest and the top of the outer slope of barrier reefs, either around high islands (Tahiti, Moorea), almost-atolls (lagoon area > emerged island area: Bora-Bora, Maupiti), or atolls, are directly impacted by oceanic high-energy swells (Guilcher, 1988). Reefs adapt to this high-energy level by (a) developing a spur- and -groove system which provides geomorphologic-hydrodynamic resistance to high-energy swells, energy absorbance, and porosity, (b) having high gross primary production and calcification rates in the algal-coral ecosystem (Hatcher, 1985), and (c) developing complex community structure as a response to highly variable environmental gradients (Fagerstrom, 1987).

Three major geomorphologic units are commonly developed in Polynesian reef systems. The first unit is a steep outer slope with continuous living algal-coral structure down to a depth of 60–80 m. The second unit is a carbonate rim dotted with fossil conglomerates upon which lie flat islands of detrital material (rubbles and sands), locally known as motu. The third unit is a lagoon of varying water depth, which varies from 0 m in a sediment-filled lagoon to 60 m in some particularly deep lagoons. Calcareous sediments line the lagoon floor, and coral pinnacles and patch reefs can also be located in the lagoon. The barrier-reef flats may be breached by passes through which lagoon waters ebb and flow. The smallest gaps in the reef flat, locally known as hoa, are shallow channels (tens of centimeters) across the reef flat through which oceanic waters can enter the lagoon.

In some reef channels, immediately below the lower limit of living corals, an impermeable apron of well-cemented carbonate sediment is found. Cementation in this environment is favored because TSW is oversaturated with respect to carbonate, especially aragonite which is five times oversaturated. This impermeable apron, which prevents horizontal exchanges between seawater and the interstitial-fluid system, is progressively dissolved below 400–500 m where seawater (AIW) becomes undersaturated with respect to aragonite; below 1 km, both aragonite and calcite are undersaturated.

Generally, the crest and reef flat of barrier and atoll reefs barely crop out at normal sea-level height. A first-order approximation based on the four archipelagoes of French Polynesia, comprising 15 high islands with barrier reefs and 80 atolls, indicates that more than 85% of the reef system is emergent, while the remainder is slightly uplifted or drowned.

The fifth archipelago, the Marquesas, constitutes an exception having only fringing reefs. After decades of contradictory statements addressing the absence of barrier reefs, there is now evidence of a drowned reef encircling each of the ten high islands of Marquesas at –95 m (Rougerie et al., 1992c.). Other drowned reefs exist in the Tuamotu Archipelago: Portland Bank, south of Gambier (almost-atoll) is now at –50 m and continues to sink (Pirazzoli, 1985); south of Niau Atoll (152°W, 15°S), a drowned atoll or guyot has been recognized at −1,000 m (Le Suavé et al., 1986).

Northeast of Tahiti, the barrier reef remains 7–15 m below sea-level for >10 km. The deleterious effect of freshwater runoff is not thought to be responsible for keeping the barrier reef from fully developing to reach the height of sea level. Passes constitute interruptions in the reef crest for evacuation of brackish-turbid lagoonal waters opposite river mouths. In north Moorea, south Maupiti and in atolls, passes are created by movement of the excess oceanic water accumulated in their lagoons by swells and reef-crest washover. Some reefs may be tilted (Tikehau south) or uplifted (Makatea, Rurutu east), by tectonic forces or hotspot activity. These elevated reef structures surrounded by living fringing reefs may be good analogs for islands and atolls of 20 ka when sea level was −125 m. Today, the integrity of Polynesian shorelines depends on their reefs which act as barriers protecting coastlines and plains from incident wave energy.

Resistance of barrier-reef rims to oceanic high energy is promoted by coral colonies and algal encrustations, as well as by early cementation that binds dead corals blocks and rubble. Early cementation is active in high-energy zones (Aissaoui and Purser, 1986). Dolomitization is another diagenetic process that increases the strength of barrier reefs, allowing them to persist for tens of million of years as in West Tuamotu Archipelago (Humbert and Dessay, 1985). Dolomite is found deep within atolls (Mururoa [q.v., Chap. 13]), at the top of atoll reefs (Tikehau) or in uplifted atolls (Makatea [q.v., Chap. 14]) and barrier reefs (Rurutu in the Cook Islands [q.v., Chap. 16]). Some Tuamotu atolls are surprisingly small; a dozen (e.g., Tepoto, Vanavana and Pinaki) have diameters of 2–5 km, giving total emergent area <20 km^2. These atolls, rising from depths of 1–3 km from adjacent ocean floor, appear like the tips of chimneys with their dense crown of coconut trees "magically" emerging over the immense blue ocean. Severe typhoon-induced slope avalanches and normal losses of organics and sediments do not seem to alter the permanence of these islands, which provides convincing documentation of the high vitality and strength of their protective barrier reefs.

Reef gaps: hoa, passes and caverns

The various types of atolls and carbonate islands of French Polynesia can be classified according to the number and importance of interconnections (passes; shallow channels or hoa) between the lagoon and ocean (Salvat, 1985). In the

Tuamotu Archipelago, a third of the 77 atolls have passes 1–10 m deep (Rangiroa, Fakarava), a third have only hoa (Takapoto, Tetiaroa), and the last third is either totally enclosed (Taiaro), filled with sediment (Nukutavake, Aki-Aki) or uplifted (Makatea, +100 m). Some lagoons of high islands (northeast Moorea, north Huahine) can also be partly or totally filled with sediments originating from the barrier-reef rims and transported into lagoon by wave surges. These sediment-filled lagoons appear as flat plains covered with dense vegetation and coconut trees, an image somewhat different from the traditional "blue lagoon" of postcards. The oceanic water passing over the barrier-reef crest by wave surge either flows across the reef flat toward the lagoon or back to the ocean via grooves and cavities. The excess water filling the lagoon is evacuated by strong currents flowing out through the passes, sometimes forming a turbid green plume that extends far into the blue ocean. Twenty-five years of visual observations, diving and measurements have shown that oceanic water entering lagoons through hoa or passes is oligotrophic. Turbulent exchange processes, such as geostrophic pumping (Nof and Middleton, 1989) or tidal jets (Wolanski et al., 1988), do not exist significantly in the permanently stratified Polynesian ocean.

Another possible evacuation process calls for percolation of hypersaline water through the porous bottom and flanks of the lagoon. This process appears to be especially important in enclosed atolls such as Taiaro or Takapoto (Rougerie, 1983), where salt exportation can reach 3.10^{-2} g m^{-2} s^{-1}. However, these lagoons are permanently more saline (38–43 psu) than the ocean because water lost by evaporation is compensated by input of oceanic water via hoa. Hoa can carry huge volumes of water during high tides and tempests. At low tide, excess lagoon water can escape only through breaks, for a certain tide range, and the volume of escaping water is a function of lagoon area. Thus, large lagoons (600–1,600 km^2) have potential outflow 50 to 100 times greater than small lagoons (2–100 km^2) if tides are equal (Bonvallot et al., 1994). In all cases, outflow is through reef-breaks, which in large lagoons, causes them to be carved and enlarged to the stage of passes (depth >1 m). Current speeds of 5–12 kn are recorded in passes of large atolls such as Rangiroa, Fakarava and Hao, and constitute a hydrodynamic force limiting coral growth and buildup in the pass channel. Accordingly, large quantities of sediment are expelled from these lagoons in strong outflow regimes. In small atolls, modest outflows cannot erode hoa to the pass stage, and absence of sediment purge favors the infilling of the lagoon (Table 15-1).

Big caverns can puncture reef slopes as in the west of Rangiroa Atoll, at 50–80 m in a zone of apparent dissolution: coral spurs do not exist there and a large population of heterotrophs, such as filter-feeders like *Stylasterina*, have colonized the reef slope. A 60-m-deep cavern with calcite stalactites has been found in the north Raiatea lagoon. The fact that this hole is not choked by surrounding sediments suggests an active circulation and/or dissolution process by interstitial reef waters. Circulation between the bottom of the lagoon of Vanavana Atoll (−5 m) and the ocean (at unknown depth) may be the result of a suction vortex that develops during low tide and sends clear oceanic water into the lagoon during rising tidal flows. This tunnel crossing beneath a 200-m-wide emergent rim may exchange water at the rate of

Table 15-1

Summary of select geomorphologic features in atolls of Tuamotu

	Number of Atolls	Lagoon Depth (m)	Surface Area (km^2) Max.	Min.	Avg.
Atolls with several passes[1]	10	> 30	1640	152	659
Atolls with one pass	17	20 ± 10	609	50	336
Atolls with no pass	44	10 ± 5	184	2	35
Atolls with filled lagoons	5	na	29	2	9

[1] passes are defined as a passage > 1 m deep across the barrief reef.
na = not applicable.

1–5 $m^3\ s^{-1}$, a sufficient flux to explain water-level variations in this quasi-enclosed and small (5 km^2) lagoon. A similar hole is known to exist in Tepoto Atoll (Tuamotu).

Lagoon waters. Lagoon waters are generally less depleted in nutrients, chlorophyll and plankton than oceanic TSW (Table 15-2). This difference is correlated with the residence time of lagoon waters and fluctuates considerably (Delesalle and Sournia, 1992). The residence time varies from weeks to years, depending on the hydrodynamic forces of the ocean, the number and depth of passes/hoa, and the size and volume of the lagoon. Short residence times reflect free exchange with the ocean and tend to maintain lagoon-water composition close to that of the intruding oligo-

Table 15-2

Summary of the hydrogeochemistry of reef interstitial waters (RIW), lagoon waters, and seawter at Tikehau Atoll[a]

Borehole[#]	Depth (m)	Salinity (psu)	N[*] (μM)	NH_4 (μM)	PO_4 (μM)	SiO_2 (μM)	pH	Redox (mV)
P_1 and P_2	1–10	25.83	2.59	2.58	1.24	4.14	7.61	8
P_1 and P_2	20–30	34.55	3.76	0.72	1.09	7.71	7.67	−60
P_3	4–17	35.87	0.23	7.15	1.08	2.24	7.61	−60
P_4 and P_5	3–11	35.86	3.48	0.59	0.49	2.80	7.79	126
P_4 and P_5	19–33	35.73	2.37	1.19	0.84	5.74	7.73	73
Lagoon	0–20	36.06	0.20	0.30	0.26	0.81	8.24	218
Ocean TSW	1–100	36.05	≤ 0.1	≤ 0.1	≤ 0.2	≤ 0.2	8.31	192
Ocean AIW	> 500	34.50	20.0	0.10	1.80	15.0	7.90	150

[a] numbers listed are average values of borehole measurements of RIW made from 1989–1992. Lagoon and seawater measurements were made from 1986–1992.
[#] P_1 and P_2 (reef crest) interstitial water is spiked by groundwater discharge from the motu phreatic lens that creates alternating oxic-anoxic conditions. P_3 (lagoon pinnacle) interstitial water is highly anoxic, except in the shallowest section facing lagoon waves. P_4 and P_5 (reef crest) have no brackish interferences and a deep oxic layer.
[*] $NO_3 + NO_2$

trophic TSW. Conversely, lagoons in enclosed or slightly uplifted atolls have long residence times, leading physico-chemical properties to shift away from ocean values, and can accumulate dissolved nutrients and particulate organic matter. However, it is important to note that this organic richness represents a shift towards natural eutrophication and tends to eliminate coral colonies to the benefit of plankton, benthic macro-algae and cyanobacterial mats. Steady-state reef-lagoon systems constitute organic oases in the oceanic desert and potential net exporters of organic and carbonate-rich matter. Such losses are balanced in the medium and long term by the net production/calcification of the barrier reef.

Motu interstitial waters. Motu composed of coral sediments and sandy gravels often occupy the shoreward/backward part of barrier and atoll reefs and can store rainfall as groundwater or in a meteoric lens that floats over the denser, saline interstitial water. This underground reservoir is filled during the rainy season but permanently discharges to the ocean and lagoon; after several dry months, as often observed in Tuamotu atolls, the groundwater may be partly withdrawn, with negative consequences for the vegetation and the life of Tuamotu population. As proposed by the Ghyben-Herzberg principle, the freshwater volume stored underground depends on two factors, the elevation of the motu above sea level and its size (Buddemeier and Oberdorfer, 1986; 1988). Atolls like Scilly or Toau have small motu and small storage capacity. Conversely, closed lagoons are totally surrounded by a continuous, (10–103 km) broad (0.3–1.5 km) and uplifted (+2 to +8 m) motu; their storage capacity is considerable with the result that groundwater leaks can permanently lower the salinity of lagoon waters. For example, lagoons of Mataiva and Niau have salinity from 32–25 psu, despite the fact that the Tuamotu is a region with a negative P-E value. The ecological consequences of this low-salinity lagoon water are important because these brackish lagoons are unfit for coral settlement but they are highly favorable to the development of macro-algae (e.g. Caulerpa) and thick cyanobacterial mats (Defarge and Trichet, 1985).

The maximum rainfall storage capacity is reached in completely filled atolls (Aki-Aki, Tikei) or in very large motu surrounding high islands (Bora-Bora; Maupiti) where underground freshwater is pumped through by under-lagoon pipes to villages located on the main basaltic island. In Amanu Atoll, the head gradient has generated sufficient brackish-water seepage to provoke the collapse of several square meters of the flanks of the pass. It is possible that such a process, by maintaining a permanent erosion of the flank of the pass, participates to the onset and long-term existence of these passages across the atoll rim (Fichez et al., 1992). Indeed, this hypothesis is consistent with the observation that for the 27 atolls with 1 or 2 passes (Table 15-1), 22 of these passes are through emergent motu, whereas the other 5 are through overflow over a reef-flat rim.

Groundwater of motu is rich in nutrients, the concentrations of which increase with depth. Vegetation like coconut trees grow remarkably on that nutrient pool and can produce 2–4 tons ha^{-1} y^{-1} of copra, without any addition of fertilizer. Motu can also have ponds or cavities where fresh groundwaters freely appear; these ponds may be flooded during high tides or tempests by lagoonal or oceanic waters, causing them

Table 15-3

Summary of the hydrogeochemistry of the brackish kopara ponds of the motu of Tikehau Atoll[a]

	Salinity (psu)	N* (μM)	NH_4 (μM)	PO_4 (μM)	SiO_2 (μM)	pH (μM)	Redox (mV)	Total Alkalinity (eq m^{-3})
Free water								
Surface	8	0.3	1	0.3	2	8.5	50	1.4
Bottom+	20	0.6	3	0.6	4	9.4	150	1.8
Interstitial Water								
5 cm	15	0.5	15	2.5	6	7.6	−200	3.5
50 cm	25	0.7	25	4.0	12	7.5	−300	2.5

[a] based on measurements made in 1991 and 1992
*NO_3 + NO_2
+ 0.5–1 m

to be brackish with salinities of 10–30 psu. These ponds are nutrient-rich both in their free water and interstitial portion (Table 15-3).

The ponds are generally colonized by algal and cyanobacterial mats named "kopara." The kopara mats, which can be >1 m thick, are highly productive and have high concentrations of chlorophyll and carotenoid pigments (Defarge and Trichet, 1985). These kopara deposits are viewed as a stromatolitic facies (MacIntyre and Marshall, 1988). In closed brackish lagoons (Niau Atoll), kopara occupy the entire area and accumulated in several distinct layers (1–6 m thick) as has been documented by subsurface drilling. Layers of fluorapatite are found inside dead kopara, in conjunction with deep anoxic conditions. In case of partial desiccation of the kopara mats, such as in the reticulated lagoon of Mataiva or the uplifted island of Makatea [q.v., Chap. 14], fluorapatite comprises thick layers, producing tens of millions of tons of ore with 30% phosphorus content. The apparent association between accumulation and degradation of dead kopara and in situ precipitation of apatite is not fortuitous, but can constitute a driving process leading to phosphogenesis (Rougerie et al., 1994). This new model of atoll phosphogenesis is important because more traditional models such as the bird-guano model, have been recently rejected for quantitative and qualitative geochemical reasons (Roe and Burnett, 1985; Bourrouilh-Le Jan, 1992; Whitehead, 1993). Indeed, the newly proposed kopara model may solve the long-standing problem of the origin of phosphate deposits at Makatea, a problem previously noted by Menard (1986).

Patch reefs and pinnacles

The abundance of corals in lagoons shows considerable variability, both in species number and in area occupied. In narrow lagoons (Tahiti, Moorea), corals are most abundant on the barrier reef and in fringing reefs. In broad lagoons of almost-atolls (Bora-Bora, Maupiti), coral settlement is mainly on the outer barrier reef and secondly as patch reefs and pinnacles, apparently scattered in a chaotic way (Guilcher, 1991). The same pattern exists in Tuamotu atolls, where some lagoons have numerous pinnacles covering up to 10% of the lagoon surface (Takapoto, Tikehau), whereas other lagoons have very few (Rangiroa, Fakarava) or none (Tetiaroa, Taenga).

These coral structures are colonized largely by varied invertebrates, especially bivalves and surrounded by a halo of fishes. Hence, lagoon biomass is correlated with the density of pinnacles. In deep lagoons, pinnacles are tall structures with steep flanks and rise from the sandy bottom to the lagoon surface. Some pinnacles may reach 50 m high, with outcropping flat tops covering 10–100 m^2, with the most productive sector facing the dominant winds and currents. In lagoonal areas lacking pinnacles or patch reefs, the bottoms are monotonous sandy plains of white sediment originating from the barrier reef: productivity of these white bottom sectors is very low, especially in shallow waters (Le Borgne et al., 1989). *Mutatis mutandis* pinnacles are to lagoons what atolls are to the ocean: highly productive stalagmitic oasis, where coral reefs develop and are surrounded by clear oligotrophic waters.

In summary, reef geomorphology can be seen to be a function of oceanic energy, water turbidity and ocean productivity (Fig. 15-3) There are four major features of the reef-atoll systems of Polynesia:

(1) The outer barrier reef is common to all of these reef systems (pure atolls, tilted atolls, uplifted or enclosed atolls, barrier reefs of high islands). This biogenic carbonate structure, which acts as a wall encircling the lagoon, is entirely built and permanently reinforced by the linked actions of primary production, calcification and early cementation that take place within the algal-coral ecosystem. Without this protective living wall, atolls and lagoons would not exist. The barrier reef is the first-order structural feature of carbonate islands, whereas lagoons range from second-order feature to being absent, as in the case of filled lagoons or uplifted atolls.

(2) Lagoonal pinnacles appear to have a chaotic distribution: abundant in some lagoons, discrete or absent in others. Much like barrier reefs, these pinnacles constitute oases for life and high productivity/calcification, compared to low productivity of lagoonal waters.

(3) Atoll enclosure and elevation control lagoon salinities, even though the Tuamotu atolls are in a zone where evaporation dominates (P-E $< - 50$ cm y^{-1}). In closed atolls with hoa, salinity can reach 43 psu with salt excess exported by water percolation through bottom and flank sediments. In closed atolls with continuous motu, freshwater stored in the phreatic lens during the rainy season can lower the lagoon salinity to <25 psu. In uplifted atolls, rainwater input leads to erosion and karstification of the carbonate structure and formation of freshwater caverns (Fig. 15-4).

(4) The spatial distribution of Indo-Pacific reefs and atolls as well as the abundance of coral colonies within lagoons confirm that corals thrive in clear oligotrophic waters and disappear in green plankton-rich waters (Hallock and Schlager, 1986; Hallock, 1988). This apparent paradox, commonly expressed as the Darwin paradox (Atkinson, 1988), needs to be solved by a convincing heuristic model. It is for that reason that, after having proposed the concept of reef functioning by geothermal

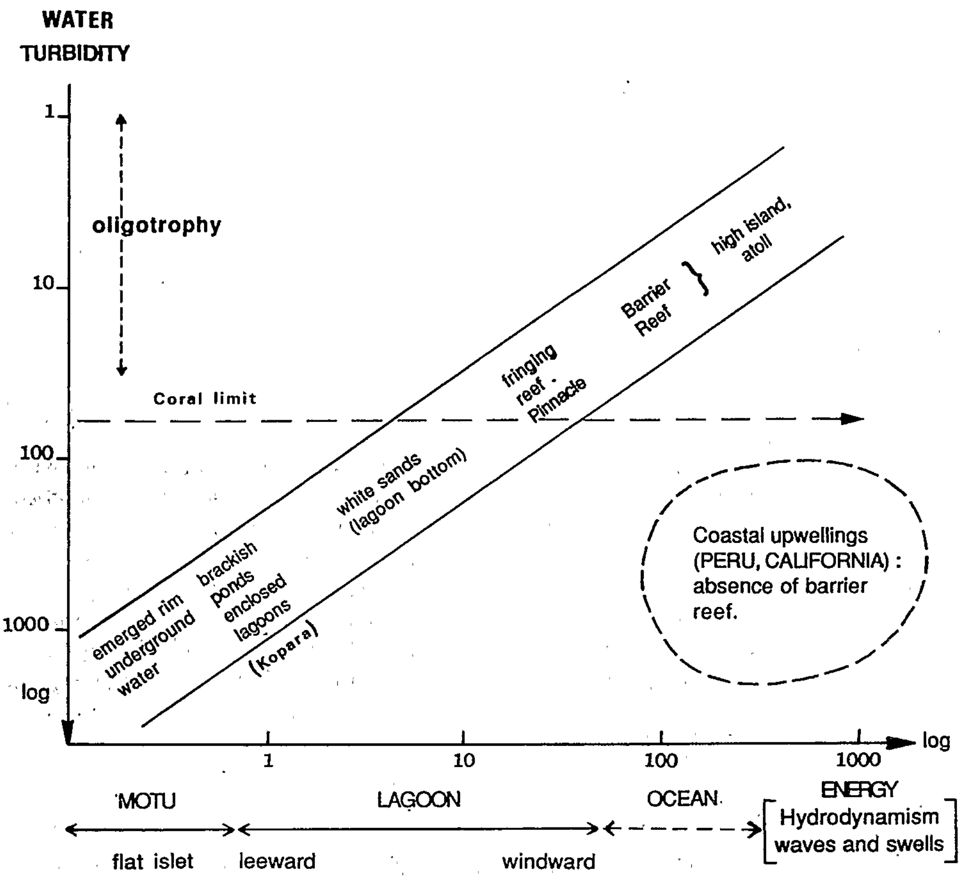

Fig. 15-3. Relation between coral-reef geomorphology and oceanographic energy regime and water turbidity (arbitrary units). Barrier and atoll reefs thrive best in coastal regions characterized by high-energy conditions and low-turbidity seawater. Barrier reefs are absent in zones of coastal upwelling. In a lagoon setting, pinnacle abundance and distribution appear chaotic. Low productivity and white sediments characterize 80–95% of the lagoon area. The water in enclosed lagoons is often hypersaline (e.g., Takapoto and Taiaro) or brackish (Niau) and coral colonies are replaced by macroalgae and/or algal mats (kopara in French Polynesia).

endo-upwelling (Rougerie and Wauthy, 1986, 1988, 1993), we have tried to maintain and support this new and controversial model by data obtained from holes drilled in atoll and barrier reefs in Polynesia.

CASE STUDY: INTERSTITIAL WATERS OF REEFS AND ENDO-UPWELLING

Previous studies of fluid flow in the subsurface of Florida and Enewetak Atoll have documented the existence of internal geothermal circulation, now often referred to as Kohout circulation, that has geologic and diagenetic consequences (Kohout, 1965; Fanning et al., 1981; Saller, 1984). The geothermal endo-upwelling concept (Rougerie and Wauthy, 1986; 1988; 1993) links thermally driven convective circu-

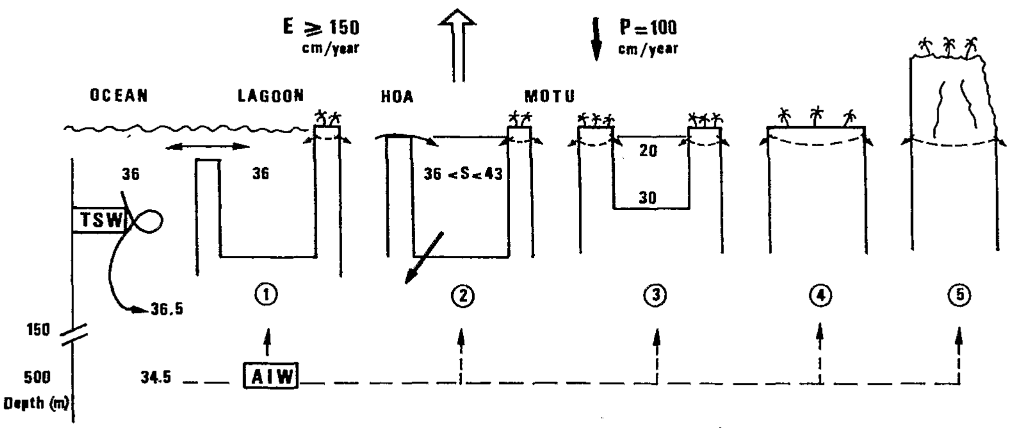

Fig. 15-4. Diversity of lagoons in Tuamotu Archipelago is a function of the amount of island enclosure and/or the elevation of motu and hence freshwater storage. Salinity values are in practical salinity units (psu). (1) Atolls with a deep oceanic pass typically have lagoon salinities that are equivalent to that of the ocean (e.g., Tikehau and Rangiroa). (2) Atolls without a deep oceanic pass, but with hoa typically have lagoon salinities that can reach 43‰ (e.g., Takapoto and Taiaro). (3) Atolls with continuous emergent motu typically have lagoon salinities that are <25‰ due to freshwater discharge from the motu to the lagoon (e.g., Niau). (4) Atolls in which the lagoon is completely filled with carbonate sediment typically have a maximum of freshwater storage (e.g., Aki-Aki and Nukutavake). (5) Uplifted atolls are subjected to erosion and karstification and may have caverns that are filled with freshwater (e.g., Makatea).

lation of subsurface fluids (i.e., Kohout circulation) with the biological consequences of this physical thermo-convective process on coral-reef growth.

The geothermal endo-upwelling process is particularly effective at carbonate islands because of the combination of a geothermal heat source and a porous and permeable geologic structure. Because of the cumulative buildup of heat, possible only in the absence of eddy diffusion, interstitial seawater within the reef framework loses density and a slow convective circulation is established: nutrient-rich deep ocean water penetrates the foundations of the island (basalt and/or carbonate) and ascends toward the top of the barrier or atoll reef where it escapes along the most permeable paths (i.e., mainly through the algal reef-crest spur- and -groove zone where sedimentation and porosity occlusion are prevented by ocean turbulence). Secondary circulation of endo-upwelled waters along sublagoonal faults and cracks on the lagoon bottom permits this water to escape and provide nutrients for the development of reef pinnacles within the lagoon. The nutrients supplied by endo-upwelled water promotes coral calcification via linkages to photo-autotrophic polyp growth. Thus, geothermally driven endo-upwelling can be considered as a necessary and sufficient process for the origin of hermatypic corals and continuing reef growth. The corollary is that algal-coral ecosystems of barrier and atoll reefs are biogeochemical signals marking the locations of interstitial water seepages. To test the validity of the concept for atolls and barrier reefs located in oligotrophic oceanic waters of the South Pacific Gyre, borings were drilled at two locations in French Polynesia in 1988, 1990 and 1992.

Tikehau Atoll reef

Reef features. Tikehau Atoll (150°W, 15°S) is located at the northwest end of the Tuamotu Archipelago (Fig. 15-5) Volcanoes that form the base of the atoll reef were probably formed 80–85 Ma, and their activity ceased in the Late Cretaceous to Early

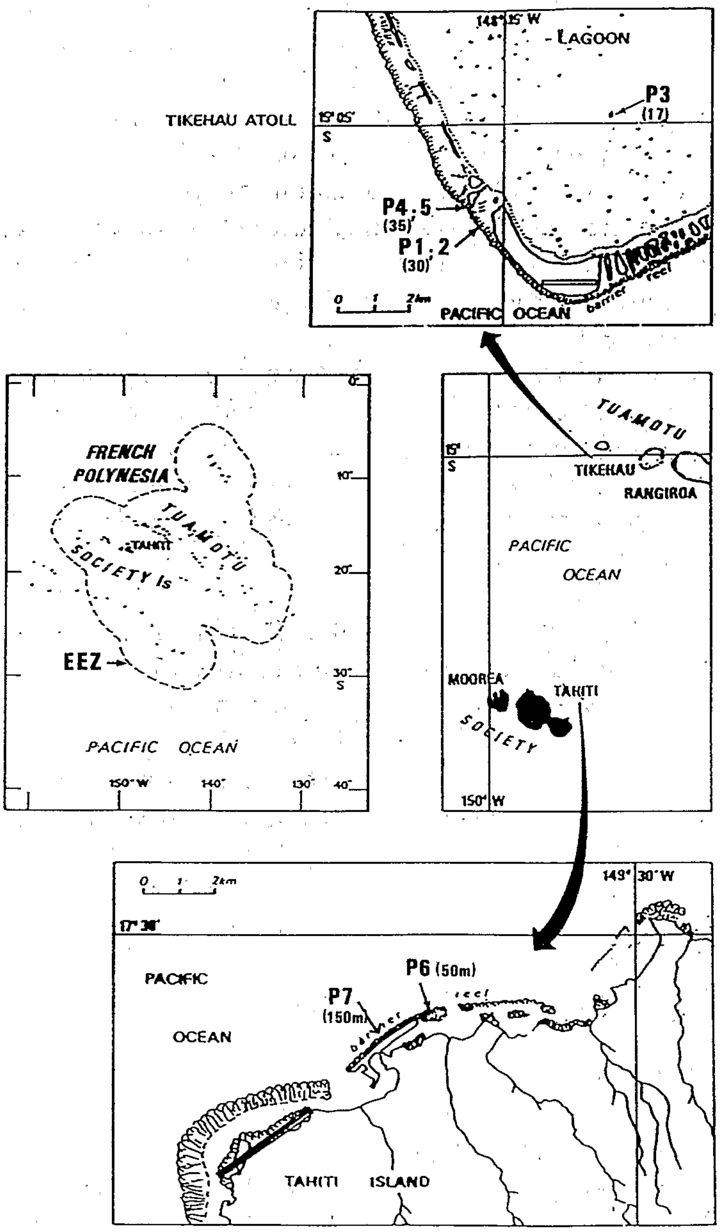

Fig. 15-5. Map of French Polynesia including the 200-miles contour outlining Exclusive Economic Zone (EEZ). Location and borehole sites for Tikehau Atoll (P_1, P_2, P_3, P_4 and P_5) and Tahiti barrier reef (P_6 and P_7). Interstitial waters are pumped at different levels via tubes that are permanently inserted inside these boreholes.

Eocene (Brousse, 1985; Montaggioni, 1993). Tikehau Atoll has a large lithified reef flat which is slightly above sea level for most of the year. High sea level due to low atmospheric pressure combined with rough seas causes waves to invade and sweep the reef-flat surface and numerous hoas. The buildup of hydraulic head in the lagoon is resisted by evaporation. Nevertheless a strong output of lagoon water occurs through the only pass, which is 4 m deep and is located on the west part of the atoll rim.

Four boreholes were drilled about 30 m from the ocean surf zone on the reef-flat and are representative of high-energy environments. The first two boreholes, denoted P_1 and P_2 (4 m apart), are also located 100 m from the emergent coral island (motu) and provided access to sampling depths of 1, 4.5, 10, 20 and 30 m. Two other boreholes, denoted P_4 and P_5 (1 m apart), were drilled on the reef flat more than 1 km away from the closest motu and provided access to sampling depths of 3, 6, 11, 19, 27 and 33 m. Another borehole, denoted P_3, was drilled through an outcropping pinnacle located in Tikehau lagoon and gave access to sampling depths of 4, 10 and 17 m. These boreholes allow comparison between interstitial waters originating from both high- and low-energy environments.

Analysis of the reef cores revealed a strongly dolomitized carbonate sequence for the whole borehole depth. Due to the hardness of dolomite, core retrieval was nearly 100% and drilling showed no large megaporosity voids. Nevertheless, cores (4 cm in diameter) had apparent porosity of 30% with mm- to cm-scale cavities. In the pinnacle, no dolomite exists, the carbonate is relatively soft, and large cavities (up to 1 m) were encountered during the drilling.

Interstitial water survey (1989–1992). Reef interstitial water (RIW) from boreholes P_4 and P_5 is oxic to a depth of 27 m and suboxic to 33 m (Table 15-2). Within boreholes P_1 and P_2, oxic conditions prevailed at the top and at the bottom of the borehole, but anoxic conditions occurred between 10 and 20 m. This anoxic layer results from the extension toward the ocean of the brackish interface between the freshwater lens (stored below the main motu) and the underlying interstitial seawater (Fig. 15-6) In borehole P_3 (pinnacle), RIW is anoxic from top to bottom. Concentrations in dissolved inorganic nutrients are significantly higher in RIW than in the surface oceanic layer. Within the boreholes, the increase in phosphate concentrations and a shift from nitrate to ammonium correlate with a shift in redox potential from positive to negative values due to biogeochemical processes and indicate a change from oxic to anoxic RIW.

The deep extension of oxic waters within the atoll reef indicates that some oxygenation processes must exist. Organic matter buried within the reef framework comes from the benthic system and undergoes mineralization with subsequent depletion in dissolved oxygen, as observed in the pinnacle borehole P_3. To compensate for this depletion, oxygen must be supplied in at least equal amount to the oxygen consumed. Our study of borehole P_3 shows that molecular diffusion of oxygen alone cannot account for this supply. Due to the observed difference between the lagoon pinnacle and the atoll RIW, it is obvious that wave surge plays a major role in providing oxygen to the interstitial system (i.e., intense wave-driven circulation in-

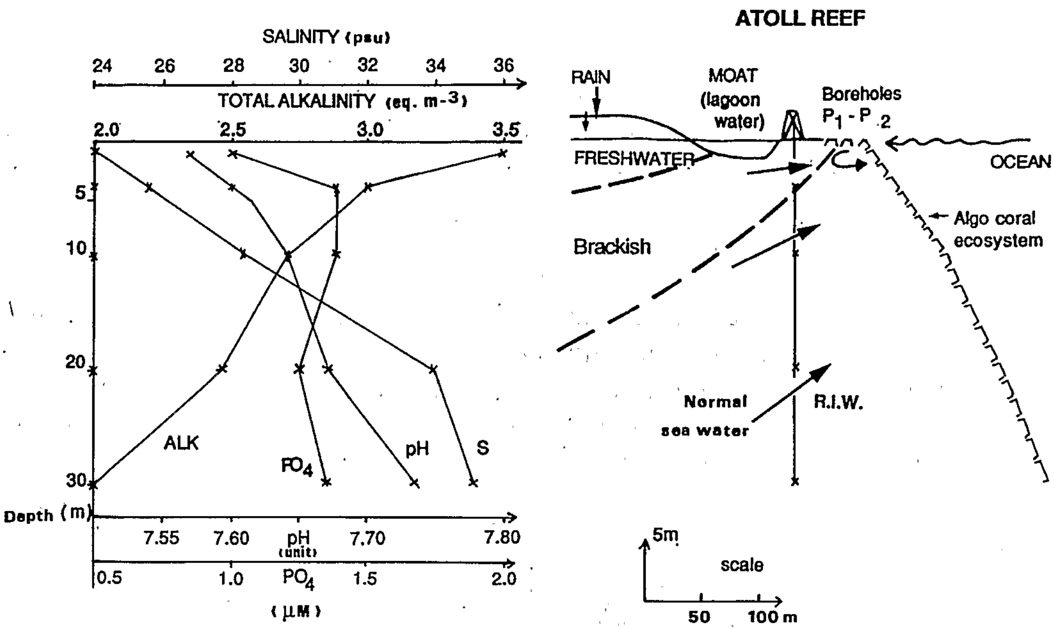

Fig. 15-6. Vertical profiles of salinity, total alkalinity, pH and inorganic dissolved phosphate in Tikehau Atoll boreholes P_1 and P_2. Groundwater discharge from the motu creates the brackish system located in the top 10 m. The low pH and high alkalinity of this brackish layer is indicative of dissolution of calcium carbonate of the reef matrix. Normal salinity (S >35 psu) seawater (RIW) is present in the boreholes by 30 m. Main seepage zone is at the reef crest, which is characterized by high energy and high porosity and hydraulic conductivity.

jects oxygen-saturated oceanic water into the reef framework, lowering the depth of the oxic-anoxic interface).

Within the oxic interstitial environment, dissolved inorganic nutrients and CO_2 are liberated in proportion to oxygen consumption. The apparent oxygen utilization (AOU) may thus be used to assess the fraction of nutrients that come from the recycling of organic matter (D'Elia, 1988). Previous calculations estimated mineralization to contribute up to 50% to the nutrient pool with the remainder originating from exogenous deep sources (Rougerie et al., 1990). This conclusion supports the geothermal endo-upwelling circulation that considers new nutrients to come from the nutrient-rich Antarctic Intermediate Water (AIW).

Salinity is a conservative parameter and provides information on the origin and the mixing of waters within the porous carbonate framework (Table 15-2). Salinity in boreholes P_1 and P_2 is used to identify a low-salinity layer at a depth of 1–10 m related to the freshwater lens of the atoll motu. At 10–20 m, salinities are 30–34 psu, values that are significantly below ocean surface salinity (36.1 $\pm$ 0.1 psu). Thus, despite being situated on the reef flat 100 m away from the island and separated from it by a shallow channel continually flushed with ocean water, boreholes P_1 and P_2 are significantly affected by freshwater intrusion from the meteoric phreatic lens. This feature agrees with recent work demonstrating the brackish transition zone to extend oceanward even when covered with a thin layer of seawater. This layered structure is due to the combined effect of freshwater flowing toward the ocean and the underlying brackish and seawater flowing upward (Moore et al., 1992; Underwood et al., 1992). The motu effect becomes undetectable below 30 m, where salinity is 35.5 psu. In boreholes P_4 and P_5, where freshwater input (meteoric or groundwater) is not suspected due to the remoteness of motu, salinities range from 35.9 psu at 6-m depth to 35.7 psu at 27 and 33 m. These salinity values are significantly lower than those of oceanic TSW. The decreasing gradient with depth agrees with the input of "endo-upwelled" AIW having a salinity of 34.5 psu at depths of 600–800 m, shown in the mixing curves between AIW, TSW and RIW (Fig. 15-7).

Strong evidence of the presence of water originating from deep-sea sources within the reef interstitial network has been gained from the study of the distribution of ^{3}He (Rougerie et al., 1991). Distribution of δ^3He in the deep Pacific shows that primordial ^{3}He is being dispersed by hydrothermal venting on the East Pacific Rise at 2 ± 0.5 km depth. The δ^3He-enriched plume spreads westward into the central

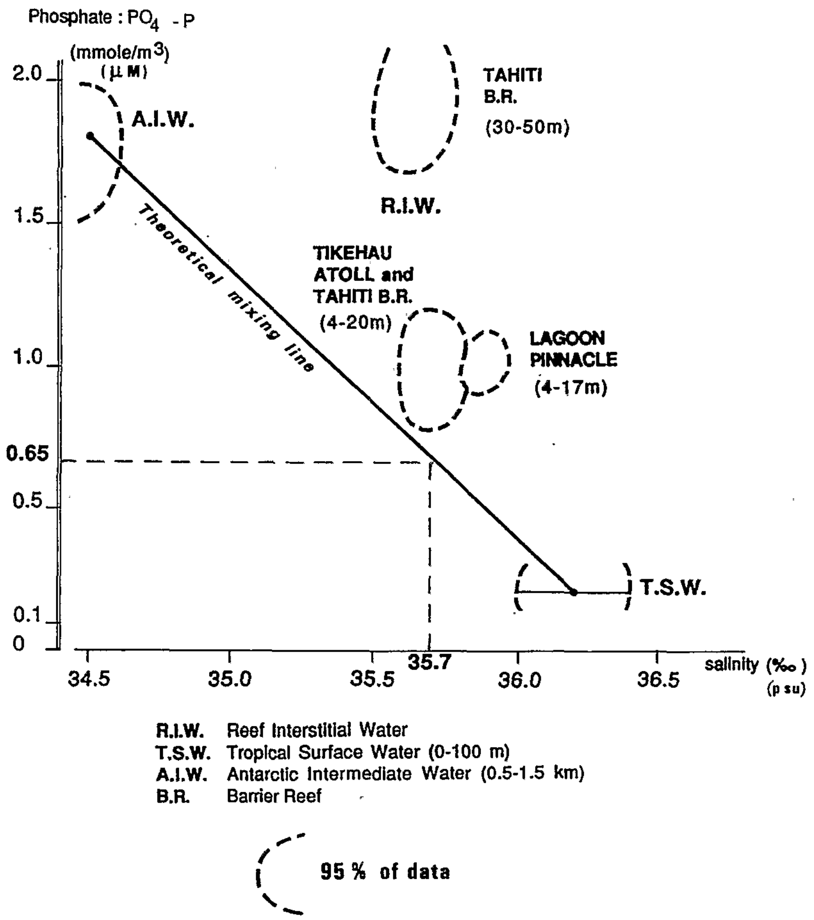

Fig. 15-7. Comparison of dissolved inorganic phosphate concentrations in oceanic (AIW and TSW) and reef interstitial waters (RIW). Phosphate concentration in RIW exceeds 0.65 mmole m^{-3}, which is the theoretical concentration of mixed AIW and TSW. This relation indicates that the chief phosphate sources are AIW plus in situ remineralization of organic matter in the reef matrix. A 0.5 psu salinity difference between RIW (35.7 ± 0.1 psu) and TSW (36.2 ± 0.2 psu) has been determined over a 3-year period (1990–1992; Rougerie et al., 1992a).

Pacific as far as the Tuamotu Archipelago where δ^3He values are up to 10% within AIW at 800-m depth. δ^3He values in Tikehau borehole waters increase with depth and are significantly higher than the values measured in the mixed layer (0–150 m) of the ocean (Fig. 15-8) Plotting δ^3He against salinity suggests that interstitial water is the result of the mixing of two endmember sources: TSW has a δ^3He of −1 to −2% and a salinity about 36.1 psu; AIW from a depth of 700–800 m has a δ^3He of 8–10% and a salinity about 34.5 psu. This result demonstrates that there is an upward flow within the reef framework driving deep oceanic water (AIW) through the carbonate pile to the top of the reef interstitial water system.

Since 1940, chlorofluorocarbon (CFC) has been anthropogenically introduced into the atmosphere through refrigerants, aerosol propellants, foams, and other products. CFCs are very useful oceanic tracers because they are conservative in seawater. The CFC (F12) concentration is homogeneous (0.8–1.0 ± 0.1 pM kg^{-1}) in the oceanic mixed layer from the surface to 200 m and sharply decreases with greater depth becoming almost undetectable in the South Pacific AIW below 400–500 m (Fig. 15-7) In Tikehau, RIW shows a F12 deficiency with concentrations around 0.2 ± 0.1 pM kg^{-1} below 10 m. Such depletion in F12 with depth can be explained either by the presence of old water trapped within the reef structure or by an input of F12-depleted ocean waters from at least 500 m. The subsurface oxygen profiles (Table 15-3; Fig. 15-6) are inconsistent with the former hypothesis. There-

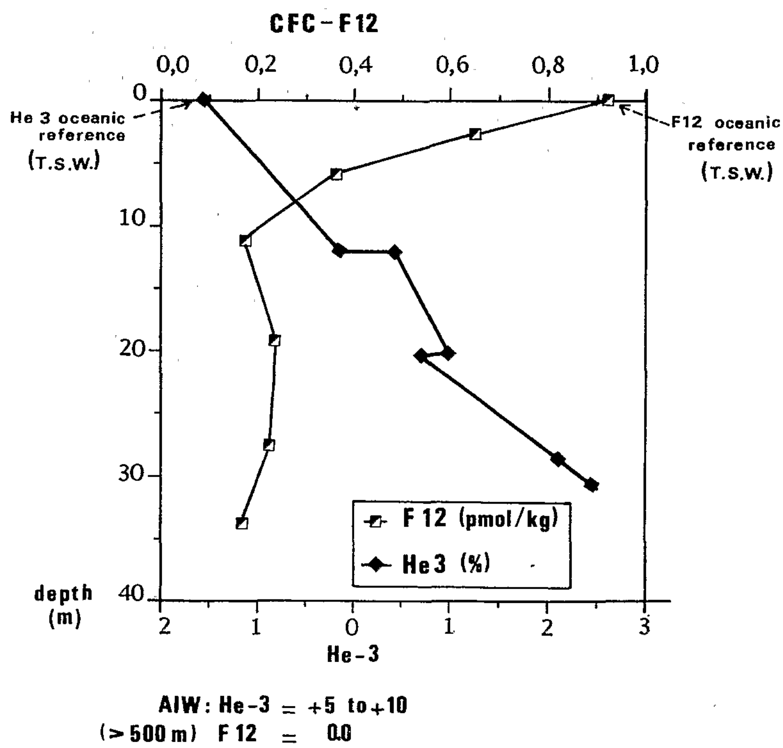

Fig. 15-8. Vertical profiles of chlorofluorocarbons (CFC-F12) and ^{3}He in RIW of Tikehau Atoll. Oceanic TSW and AIW reference values are given. Strong anomalies in the distribution of these conservative tracers into RIW can be explained by upward circulation of AIW inside reef matrix, as proposed by the geothermal endo-upwelling model (Rougerie et al., 1991).

fore, the F12 distribution strongly supports the conclusion from the study of ^3He distribution that AIW is a significant component of RIW (Fig. 15-8)

Dissolved non-aromatic hydrocarbons and fatty-acid concentrations were generally lower in the ocean than in the RIW of the Tikehau boreholes, where they increased with increasing depth (Andrié et al., 1992). Below 5 m the n-alkane profiles point to significant early diagenetic alterations due both to bacterial activity and to thermal maturation of organic matter (Bouloubassi et al., 1992). Such processes may have occurred in the deeper framework of the reef because of geothermal activity over geologic time. The presence of such mature markers in the top 30 m of the reef strongly suggests that waters follow an ascending movement from near the volcanic basement to the top. This suggests that ascending interstitial water, initially rich in dissolved organic matter from AIW and from leaching of organic matter trapped within the carbonate framework, undergoes sufficient heating in anoxic environment to produce mature alkanes.

Tahiti barrier reef (150° W, 17°30′S)

Reef features. There are a few studies dealing with the geology of carbonate reefs from high islands in French Polynesia. Boreholes have been drilled through the fringing and patch reefs surrounding Papeete harbor (Deneufbourg, 1971). However, materials from these boreholes were studied mainly from a sedimentologic perspective. Later, a 24-m-deep borehole drilled through the same reef system yielded information on sea-level variations since 7.0 ka (Pirazzoli and Montaggioni, 1986). Other drillings through a carbonate platform in Moorea Island were used to address paleohydrology issues (Faissolle, 1988).

Borehole P_6 was drilled in 1990 to a depth of 50 m through the barrier reef protecting Tahiti harbor (Fig. 15-5) Sampling tubes gave access to sampling depths of 1, 5, 20, 30 and 50 m. Unlike the reef of Tikehau Atoll, the Tahiti barrier reef lies a few tens of centimeters below sea level, is permanently flushed by waves and is emergent only in anomalous low sea levels common during peak ENSO events. The Tahiti core was studied for its petrography and mineralogy (Déjardin, 1991). Core recovery was 25–95%, with megaporosity voids (indicated by the drilling-rate logs) accounting for the low-recovery zones. Examination of the core material yielded no evidence of freshwater diagenesis, thus indicating no recent subaerial exposure events for the top 50 m of the reef. Radiocarbon dating (Bard et al., 1993) yielded ages of 3,000 and 5,500 y B.P. at depths of 2 and 3 m, respectively, corresponding to a period of relative sea-level stability. Ages regularly decreased with depth to 10,000 y B.P. at 50 m; this trend is interpreted as the consequence of a period of rapid vertical buildup of the reef in response to the Holocene eustatic sea-level rise. Today, that barrier reef is cut by two passes located in the axes of two valleys with permanent rivers (current of 0.5–2 $m^3\ s^{-1}$ with flood current $> 10\ m^3\ s^{-1}$ during typhoons). The river waters lower the salinity of the lagoon from 35 to 25 psu in the extreme case; the lagoon head, enhanced by overflow of oceanic water above the reef crest, creates current, which can reach several knots at the pass sill (10 ± 2 m) during ebb.

At the end of 1992, borehole P_7 was drilled to 150-m depth on the barrier-reef crest, 1 km west of borehole P_6. Analysis of the borehole P_7 showed the base of the reef carbonate at 110 m, followed by 30 m of mixed carbonate-volcanic detrital material (at 110–140 m) and a 10-m-thick layer of basalt (at 140–150 m). The drilling-rate log demonstrated the presence of large megaporosity voids (m^3 to tens of m^3) in agreement with observations on borehole P_6. Detailed study of the core and interstitial waters from borehole P_7 is in progress.

Interstitial water survey (1990–1992). Physico-chemical parameters (Table 15-4) for Tahiti borehole P_6 showed positive values of redox potential in the first 20 m together with the presence of free oxygen. Physico-chemical determinations confirms the turbulent penetration of aerated surface-ocean water through the outer margin of the reef, consistent with our interpretations for the reef of Tikehau Atoll. Oxic conditions sharply disappear below 20 m, demonstrating that AOU exceeds the rate of oxygen renewal. Values of pH in RIW decrease with depth, from 7.9 at the surface to 7.6 at 50 m, and are always significantly lower than those from the adjacent oceanic waters (8.3). These changes in pH values imply a correlative shift in chemical equilibrium from carbonate to bicarbonate with possible dissolution of the carbonate framework, especially within the anoxic zone.

Nitrate is the dominant inorganic nitrogenous form in the oxic zone where ammonium concentrations are low (1 μM or less). From 30-m depth, reducing conditions result in the disappearance of oxidized N species, a large increase in ammonium (up to 10 μM), an increase in phosphate (up to 2.5 μM) and a large excess in silicate (up to 80 μM). Two distinct fields of data emerge from the Tahiti borehole P_6 dataset. The first cluster contains slightly enriched values in phosphate, nitrate and

Table 15-4

Summary of the hydrogeochemistry of reef interstial waters (RIW) at Tahiti[a]

Borehole[#]	Depth (m)	Salinity (psu)	N^* (μM)	NH_4 (μM)	PO_4 (μM)	SiO_2 (μM)	pH	Redox (mV)
P_6 (reef crest)	1	35.80	2.82	1.63	0.71	17.14	7.86	211
		(0.16)	(1.48)	(0.75)	(0.20)	(6.90)	(0.17)	(26)
	5	35.71	1.63	1.67	0.91	21.27	7.78	153
		(0.13)	(1.32)	(1.02)	(0.34)	(6.64)	(0.17)	(70)
	20	35.73	1.52	0.76	1.06	21.21	7.78	111
		(0.12)	(1.22)	(0.72)	(0.60)	(5.79)	(0.16)	(88)
	30	35.78	0.16	12.00	1.56	63.62	7.65	−130
		(0.07)	(0.06)	(3.77)	(0.39)	(9.40)	(0.12)	(35)
	50	35.74	0.09	10.70	2.14	79.97	7.67	−120
		(0.11)	(0.06)	(3.97)	(0.54)	(8.11)	(0.12)	(28)

[a] numbers listed are average values of borehole measurements of RIW made from 1989–1992. Lagoon and seawater measurements were made from 1986–1992. Numbers listed in parentheses are standard deviation values.

$^*NO_3 + NO_2$

silicate relative to surface-ocean values and represents oxic waters from the top 20-m layer. The second cluster contains even higher values of phosphate, ammonium and especially silicate and represents anoxic waters from the lower 30-and 50-m layers. Such a distribution clearly indicates that excess silica is provided by an exogenous source and adds to organic-matter recycling and upward transport of AIW. Leaching of the basalt, which is composed of up to 50% of soluble silica, by interstitial water flow is likely responsible for the observed excess silicate. The higher silicate concentrations in Tahiti relative to those observed in Tikehau RIW result from differences in the depth of the carbonate-basalt contact, which is located at 110–130 m at Tahiti and is estimated to be at least 1,000 m below the flanks of Tikehau Atoll.

Salinity in Tahiti borehole P_6 (35.7 ± 0.1 psu) is lower than in TSW (36.1 ± 0.1 psu). As in Tikehau, this difference may be explained by mixing between two oceanic water sources: AIW (34.5 psu) and TSW (36.1 psu). The higher salinity range in the Tahiti borehole may reflect a higher input of TSW within the reef matrix, due either to stronger wave-surge dynamics or higher carbonate porosity.

The Tahitian RIW shows a noticeable F12 deficiency with concentrations around 0.8 ± 0.1 pM kg^{-1} at depths of 1–20 m and around 0.5 ± 0.1 pM kg^{-1} below a depth of 30 m (Andrié et al., 1992). The depletion of F12 with depth can be explained by the input of F12-depleted waters from 300–400 m, a level where oceanic values correspond with RIW values and which is thought to correspond to the base of the carbonate pile overlying the volcanic basement. The higher F12 concentrations observed in Tahiti relative to those observed in Tikehau RIW can be explained similarly to the salinity differences between these boreholes: greater mixing with CFC-rich TSW (0–150 m) or by a reduced flux through the basalts. The latter perhaps is in response to the lower hydraulic conductivity of the basalt compared to that of the carbonate sequence (Guille et al., 1993). Small variability in the tracer records probably results from heterogeneity in the reef structure, producing discontinuities in RIW circulation.

Synthesis and significance

Although the initial drillings were done to test the validity of the endo-upwelling model, study of RIW allows us to address other fundamental questions regarding the functioning of the entire atoll-reef system. The following is a synthesis of our observations:

(1) High concentrations of nutrients and carbon dioxide (CO_2) within the top of the reef matrix can support huge gross productivity within the reef system, despite the oligotrophy of the surrounding ocean. Losses of organic matter and exportation of sediment from the nutrient-rich reef to the nutrient-poor ocean can be compensated for by the net productivity of the algal-coral ecosystem. Internal upward circulation from nutrient-rich oceanic AIW to the reef crest is supported by results from studies of conservative markers such as ^3He and CFC. The Darwin paradox

(i.e., oasis of barrier reef productivity in the desert of an oligotrophic tropical ocean) can then be solved in a rational way.

(2) The distribution and vertical gradients of nutrients, CO_2 and O_2 indicate that RIW can reach anoxia (i.e., it can have intermediate to high AOU values). These results are in agreement with similar approaches developed in coastal upwelling areas. The difference between upwelling and endo-upwelling lies in the driving force; upwelling is a wind-driven process whereas endo-upwelling is a geothermally driven process. Upwelling intensity and occurrence is linked to wind-current variability; endo-upwelling depends on the local heat flow and the hydraulic conductivity and porosity of the structure.

(3) Interstitial water systems of barrier and atoll reefs contain oxic water to depths of 20–30 m, a pattern evidently dependent on the oceanic hydrodynamic forcing. This feature is of paramount importance for coral growth, organic matter recycling, and diagenesis of the carbonate framework. Oxygenation of the upper interstitial water appears to result from the mixing of CO_2-rich (low pH), anoxic deep interstitial water with CO_2-poor (high pH), oxic oceanic water injected into the reef matrix by wave surge. We propose the principle of maximum (early) cementation (Aissaoui and Purser, 1986) to be a diagenetic process linked closely to the specific state of the CO_2-carbonate equilibrium of RIW. In response to rapid CO_2 degassing at the top of the reef, this equilibrium shifts toward carbonate saturation that favors early cementation.

(4) Most pinnacle interstitial waters are anoxic and nutrient-rich and are consistent with other studies in lagoon patch reefs (Sansone et al., 1988; Tribble et al., 1990). For large, emergent, lagoon pinnacles, algal-coral growth is favored in the windward side; in contrast, ecosystem development is impaired by excess sedimentation on the leeward side. Pinnacles can be viewed as localized constructions built by corals in zones of RIW seepages. Interstitial sublagoonal circulation requires that bottom sediments in the lagoon must be crossed by faults or cracks. These coral constructions are, therefore, likely related to antecedent karst topography and are the expression of an internal hydrogeologic flow pattern.

(5) Groundwater accumulated in reef-flat islets (motu) during the rainy season escapes continuously towards the lagoon and ocean. Boreholes P_1 and P_2 have been used to monitor this outflow which shifts RIW salinity to values as low as 20–30‰ psu in the top 10 m (Fig. 15-6) This brackish water has a low pH and high alkalinity which indicates that it has the potential to dissolve reef matrix and enhance porosity. The meteoric phreatic water is vital to vegetation whose outstanding productivity is forced by the interstitial nutrient reservoir present in the whole atoll-reef structure. Discharge of fresh to brackish groundwater to the reef crest, important in the rainy season, does not alter coral-reef development (e.g., coral density or spur- and -groove patterns), but can weaken motu and the atoll rim, initiating hoa and pass development. Passes constitute, for the living ecosystem, breaches that cannot be closed when the escaping volume of lagoon water is significant, as in large atolls or when it has low salinity, as in the lagoons of high islands.

(6) Some motu have brackish ponds in locations where groundwater accumulates. These ponds are colonized by cyanobacterial algal mats, kopara. In totally enclosed

atolls with a broad and continuous motu, the volume of groundwater stored may be equivalent to or greater than the lagoon water volume. Leakage of freshwater toward the lagoon transforms it to a brackish system colonized only by thick mats of kopara, as is found at Niau Atoll. Because layers of precipitated fluorapatite occur in the internal anoxic basement of dead kopara (Trichet and Fikri, 1993), we believe this stromatolitic facies (Defarge et al., 1993) is a step in atoll phosphogenesis. Previously, Rougerie and Wauthy (1989) suggested that atoll phosphogenesis is a consequence of endo-upwelling with subsequent accumulation of phosphorus in closed lagoons, massive phosphate precipitation, and deposits as observed in sediment-filled or uplifted atolls of Mataiva, Makatea, Nauru (Bernat et al., 1991). Our data on kopara ponds show that phosphorus can be sequestered in these anoxic organic mats until the final step, which is the oxidation of these mats and fluorapatite precipitation upon emergence of the atoll (Rougerie et al., 1997).

(7) Dolomite is present in numerous reefs and atolls, sometimes at great depth. Its origin is highly controversial, but several authors have clearly linked dolomitization to thermo-convection of deep oceanic water within the porous and permeable carbonate structure (Fanning et al., 1981; Saller, 1984; Aharon et al., 1987). Recent studies of the Bahamas Banks show the efficiency of the internal circulation to perform secondary dolomitization (Whitaker and Smart, 1990). Because geothermal endo-upwelling is a thermo-convective process, we believe it has good potential in dolomitization; magnesium-rich AIW, warmed by heat flow, dissolves calcite, furnishes magnesium to dolomite crystals and the exchanged calcium evacuates upward. In some atolls fluorapatite is in direct contact with massive dolomites.

CONCLUDING REMARKS

The large geomorphological diversity of Polynesian barrier and atoll reefs can be accommodated by a single heuristic model that we call geothermal endo-upwelling. The model is based on the circulation of interstitial water driven by thermal convection and modulated at the reef surface by oceanic wave surge and secondarily by the circulation of recharge-driven meteoric water. Our geothermal endo-upwelling model, which can be viewed as a form of low-energy hydrothermalism, impacts on a diversity of biogeochemical processes including (1) the productivity, calcification and cementation processes active in algal-coral reef ecosystems, (2) carbonate and phosphate diagenesis, and (3) degradation of organic matter (Fig. 15-9) A barrier reef is not only an accumulation of dead corals and carbonate sediments topped by a living veneer of algae and corals, but a complex and integrated macrocosm in which interstitial circulation is the key factor whose involvement ranges from short-term coral growth to long-term atoll evolution. We investigated the Darwinian paradox (i.e., oasis of barrier reef productivity in the desert of an oligotrophic tropical ocean) using interstitial-water studies. The results of our investigations have led us to propose a new paradigm for the development and maintenance of the entire Polynesian reef system. More studies are necessary to evaluate the robustness

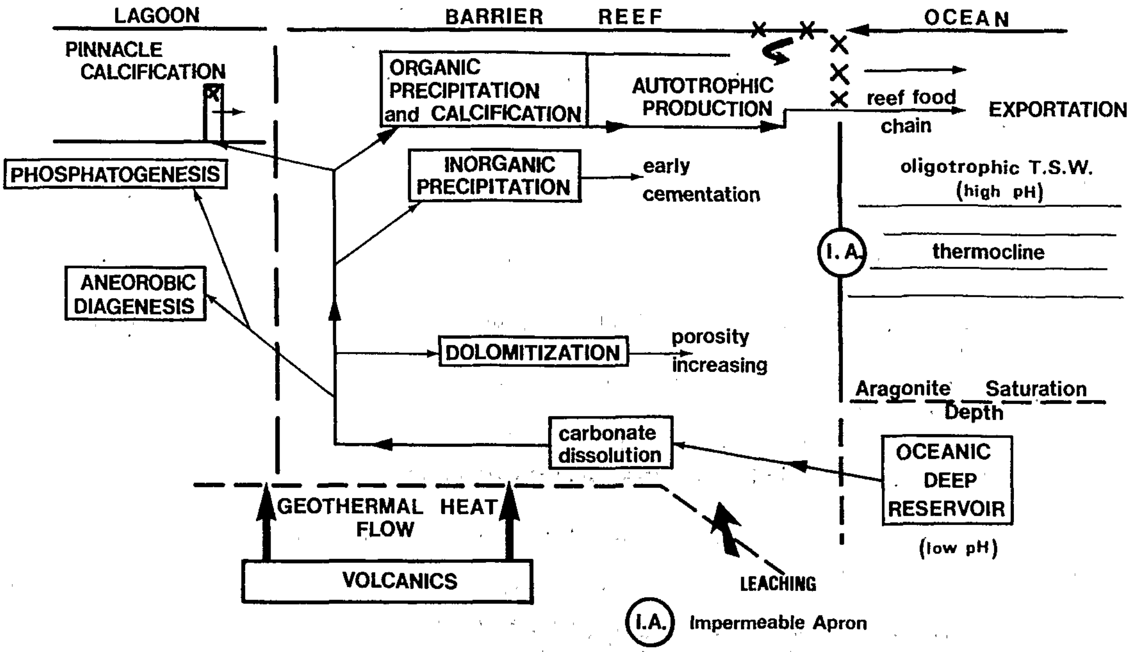

Fig. 15-9. Schematic diagram of the geothermal endo-upwelling model showing the zones of active inorganic and organic precipitation and dissolution. Flow dynamics and kinetics of the chemical exchanges are a function of heat flow, porosity, hydraulic conductivity and energy regime at the reef crest. Cementation of the impermeable apron (IA), which prevents horizontal exchange between seawater and interstitial reef water, is controlled by the carbonate saturation state of the Polynesian ocean, which is oversaturated with respect to aragonite to a depth of 400–500 m.

of our model and whether it can be applied more generally to others reef atoll provinces.

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REFERENCES

Aharon, P., Socki, R. and Chan L., 1987. Dolomitization of atolls by sea water convection flow: test of a hypothesis at Niue. South Pacific. J. Geol., 95: 187–203.

Aissaoui, D.M. and Purser, B.H., 1986. La cimentation dans les récifs: principe de cimentation maximale. Compt. Rend. Acad. Sci., 303, II: 301–303.

Andrews, J.C. and Pickard, G.L., 1990. The physical oceanography of coral reef systems. In: Z. Dubinsky (Editor), Ecosystems of the World, 25, Coral Reefs-II: 11–48.

Andrié, C., Bouloubassi, I., Cornu, H., Fichez, R., Pierre, C. and Rougerie, F., 1992. Chemical and tracer studies in coral reef interstitial waters (French Polynesia): implication for endo-upwelling circulation. Proc. Seventh Int. Coral Reef Symp. (Guam), 2: 1165–1173.

Atkinson, M.J., 1988. Are coral reefs nutrient limited? Proc. Sixth Int. Coral Reef Symp. (Townsville), 1: 157–166.

Barber, R.T., 1992. Geologic and climatic time scales of nutrient variability. In: P.G. Falkowski (Editor), Primary Productivity and Biogeochemical Cycles in the Sea. Plenum Press, New York, 89–106.

Bard, E., Montaggioni, L., Arnold, M. and Rougerie, F., 1993. C14 dating of a 50 m core from the Tahiti Barrier Reef. (Abstr.) Intern. Workshop on Intraplate Volcanism, Tahiti.

Bernat, M., Loubet M. and Baumer A., 1991. Sur l'origine des phosphates de l'atoll de Nauru. Oceanol. Acta, 14: 325–331.

Bonvallot, J., Laboute, P., Rougerie, F. and Vigneron, E., 1994. Les atolls des Tuamotu. Eds. ORSTOM Paris, 296 pp.

Bouloubassi, I., Saliot, A., Rougerie, F. and Trichet, J. 1992. Hydrocarbon geochemistry in coral reefs pore waters, French Polynesia, Proc. Water Rock Interaction, Balkema Rotterdam, 271–274.

Bourrouilh Le Jan, F., 1992. Evolution des karsts oceaniens (karsts, bauxites, phosphates). Karstologia, 19: 31–50.

Brousse, R., 1985. The age of the islands in the Pacific Ocean: volcanism and coral reef build up. Proc. Fifth Int. Coral Reef Symp. (Manila), 6: 389–400.

Brown, B., 1990. Coral bleaching. Coral Reefs, 8: 153–232.

Buddemeier, R.W. and Oberdorfer, J.A., 1986. Internal hydrology and geochemistry of coral reefs and atoll islands: keys to diagenetic variations. In: J.H. Schroeder and B.H. Purser (Editors), Reef Diagenesis. Springer-Verlag, Berlin, pp. 91–111.

Buddemeier, R.W. and Oberdorfer, J.A., 1988. Hydrogeology and hydrodynamics of coral reef pore waters. Proc. Sixth Int. Coral Reef Symp. (Townsville), 2: 485–490.

Defarge, C. and Trichet J., 1985. First data on the biogeochemistry of kopara deposits from Rangiroa Atoll. Proc. Fifth Int. Coral Reef Symp. (Tahiti), 3: 365–370.

Defarge, C., Trichet, J., Sansone, F., Tribble, J., Robert, M. and Jaunet, A.M., 1993. Nouvelles preuves de l'intervention de réseaux organiques hérités de procaryotes dans la micro-structuration et la carbonatation des stromatolites actuels. Compt. Rend. Acad. Sci., 316, II: 1107–1114.

Déjardin, P., 1991. Forage du récif barrière nord de Tahiti. Caractérisation petrographique et études hydrogeochimique. UFP Tahiti, 38 pp. + annexes.

Delcroix, T. and Henin, C., 1991. Seasonal and interannual variations of sea surface salinity in the tropical Pacific Ocean. J. Geophys. Res., 98: 22, 135–22, 150.

Delesalle, B. and Sournia, A., 1992. Residence time of water and phytoplankton biomass in coral reef lagoons. Cont. Shelf Res., 12: 939–949.

D'Elia, C., 1988. The cycling of essential elements in coral reefs. In: Pomeroy and Alberts (Editors), Concepts of Ecosystem Ecology. New York Ecological Studies, 67, Springer-Verlag, New York, pp. 195–204.

Deneufbourg, G., 1971. Etude géologique du Port de Papeete-Tahiti. Cah. Pac., 12 and 13.

Fagerstrom, A., 1987. The evolution of reef communities. John Wiley, New York, 600 pp.

Faissolle, F., 1988. Hydrogéologie, Paléohydrogéologie et diagénèse d'un système aquifère carbonaté récifal côtier. Thèse, Université Bordeaux III, 269 pp.

Fanning, K., Byrne, R., Breland, J., Betzer, P., Moore, W. and Elsinger, R., 1981. Geothermal springs of the west Florida Continental Shelf: evidence for dolomitization and radionuclide enrichment. Earth Planet. Sci. Lett., 52: 345–354.

Fichez, R., Buestel, D. and Quessu, D., 1992. Etude du phénomène de résurgence de Novembre 1991 dans la passe de l'atoll d'Amanu (Tuamotu). Archives d'Oceanogr., ORSTOM Tahiti, 11 pp.

Glynn, P.W., 1990. Coral mortality and disturbances to coral reefs in the tropical eastern Pacific. In: P.W. Glynn (Editor), Global Ecological Consequences of the 1982–83 El Nino Southern Oscillation. Elsevier Oceanogr., Ser. 52, Amsterdam, 55–126.

Glynn P.W., 1993. Coral reef bleaching: ecological perspectives. Coral Reefs, 12: 1–17.

Guilcher, A., 1988. Coral reef geomorphology. John Wiley, Chichester, 228 pp.

Guilcher, A., 1991. Progress and problems in knowledge of coral lagoon topography and its origin in the South Pacific by way of pinnacle study. In: R.H. Osborne (Editor), From Shoreline to Abyss: Contributions in Marine Geology in Honor of Francis Parker Shepard. Soc. Econ. Paleont. Mineral., Spec. Publ. 46: 173–188.

Guille G., Goutière G. and Sornein, J.F., 1993. Les atolls de Mururoa et de Fangataufa (Polynésie Française). Eds CEA/DIRCEN – GAP, 168 pp.

Hallock, P., 1988. The role of nutrient availability in bioerosion: consequences to carbonate build ups. Palaeogeogr. Palaeoclimat. Palaeoecol., 63, 275–291.

Hallock, P. and Schlager W., 1986. Nutrient excess and the demise of coral reefs and carbonate platforms. Palaios, 1: 389–398.

Hatcher, A.I., 1985. The relationship between coral reef structure and nitrogen dynamics. Proc. Fifth Int. Coral Reef Symp. (Tahiti), 3: 407–413.

Heywood K.J., Barton E.D. and Simpson J.H., 1990. The effects of flow disturbance by an oceanic island. J. Mar. Res., 48: 55–73.

Humbert, L. and Dessay J., 1985. Aspects de la dolomitisation de l'île de Makatea (Polynésie Française). Proc. Fifth Int. Coral Reef Symp. (Tahiti), 3: 271–276.

Jouannic, C. and Thompson, R.M., 1983. Bibliography of geology and geophysics of the South Pacific. UN-ESCAP, CCOP/SOPAC. Techn. Bull. 5, 258 pp.

Kohout, F.A., 1965. A hypothesis concerning cyclic flow of salt water related to geothermal heating in the Floridan aquifer. Trans. New York Acad. Sci., Series 2, 28: 249–271.

Laboute, P., 1985. Evaluation of damage done by the cyclones of 1982–1983 to the outer slopes of the Tikehau and Takapoto Atolls. Proc. Fifth Int. Coral Reef Symp. (Tahiti), 3: 323–329.

Le Borgne, R., Blanchot, J. and Charpy, L., 1989. Zooplankton of Tikehau Atoll (Tuamotu Archipelago) and its relationship to particulate matter. Mar. Biol. 102: 341–353.

Le Suavé, R., Pautot, G., Hoffert, M., Monti, S., Morel, Y. and Pichocki, C., 1986. Cadre géologique de concrétions poly-métalliques cobaltifères sous-marines dans l'archipel des Tuamotu. Compt. Rend. Acad. Sci., 303, II: 11, 1013–1018.

Levitus, S., 1982. Climatological atlas of the world ocean. NOAA Prof. Paper. US. Govt. Print. Off. Washington, D.C., 13, 173 pp.

MacIntyre, I. and Marshall, J., 1988. Submarine lithification in coral reefs: some facts and misconceptions. Proc. Sixth Int. Coral Reef Symp. (Townsville), 1: 263–272.

Menard, H.W., 1986. Islands. Freeman, New York, 230 pp.

Montaggioni, L., 1993. Volcano-isostatic polyphase uplift: a key to the post-oligocene evolution of the northwestern Tuamotu atolls (Central Pacific). (Abstr.) Intern. Workshop on Intraplate Volcanism, Tahiti.

Moore, P., Reddy, K. and Graetz, D., 1992. Nutrient transformations in sediments as influenced by oxygen supply. J. Environ. Qual., 21(3): 387–393.

Nof, D. and Middleton, J., 1989. Geostrophic pumping inflows and upwelling in barrier reefs. J. Phys. Oceanogr., 19: 874.

Pernetta, J.C. and Hughes, P.J., 1990. Implications of expected climate changes in the South Pacific region: an overview. UNEP, Regional Seas Rep. and Stud., 128, 279 pp.

Pirazzoli, P.A., 1985. Bathymetric mapping of coral reefs and atolls from satellite. Proc. Fifth Int. Coral Reef Symp. (Tahiti), 6: 539–544.

Rancher, J. and Rougerie, F., 1993. Hydropol. Situations océaniques du Pacifique Central Sud. Editions SMSR Montlhéry, 91 pp.

Roe, K.K. and Burnett, W.C., 1985. Uranium geochemistry and dating of Pacific island apatite. Geochim. Cosmochim. Acta, 49: 1581–1592.

Rougerie, F., 1983. Nouvelles données sur le fonctionnement interne des lagons d'atoll. Compt. Rend. Acad. Sci., 297, II: 909–912.
Rougerie, F. and Wauthy, B., 1986. Le concept d'endo-upwelling dans le fonctionnement des atolls-oasis. Oceanolog. Acta, 9: 133–148.
Rougerie, F. and Wauthy, B., 1988. The endo-upwelling concept: a new paradigm for solving an old paradox. Proc. Fifth Int. Coral Reef Symp. (Tahiti), 3: 21–26.
Rougerie, F. and Wauthy, B., 1989. Une nouvelle hypothèse sur la genèse des phosphates d'atolls: le rôle du processus d'endo-upwelling. Compt. Rend. Acad. Sci., 308, II: 1043–1047.
Rougerie, F. and Wauthy, B., 1993. The endo-upwelling concept: from geothermal convection to reef construction. Coral Reefs, 12: 19–30.
Rougerie, F and Rancher, J., 1994. The Polynesian South Ocean: features and circulation. Marine Pollution Bulletin 29 (1–3): 14–25.
Rougerie, F., Wauthy, B. and Andrié, C., 1990. Geothermal endo-upwelling model testing for atoll and high island barrier reef. Proc. Intern. Workshop, Nouméa, pp. 197–202.
Rougerie, F., Andrié, C. and Jean-Baptiste, P., 1991. Helium-3 inside atoll barrier reef interstitial water: a clue for geothermal endo-upwelling. Geophys. Res. Lett., 18: 109–112.
Rougerie, F., Fagerstrom, J., and Andrié C., 1992a. Geothermal endo-upwelling: a solution to the reef nutrient paradox. Cont. Shelf Res., 12: 785–798.
Rougerie, F., Salvat, B., Tatarata, M., 1992b. La mort blanche des coraux. La Recherche, 23: 826–834.
Rougerie, F., Wauthy, B. and Rancher, J., 1992c. Le récif barrière ennoyé des Iles Marquises et l'effet d'île par endo-upwelling. Compt. Rend. Acad. Sci., 315, II: 677–682.
Rougerie, F., Jehl, C. and Trichet, J., 1994. Phosphorus pathway in atoll. AGU-ASLO Meeting. La Jolla (poster).
Rougerie, F., Jehl, C., Trichet, J., 1997 Phosphorus pathway in atolls: endo-upwelling input, cyanobacterial accumulation and carbonate fluoro apatite (CFA) precipitation-Marine Geology.
Saller, A., 1984. Petrologic and geochemical constraints on the origin of subsurface dolomite, Enewetak Atoll: an example of dolomitization by normal sea water. Geology, 12: 217–220.
Salvat, B., 1985. An integrated (geomorphological and economical) classification of French Polynesian atolls. Proc. Fifth Int. Coral Reef Symp. (Tahiti), 2: 337.
Sansone, F.J., Andrews, C., Buddemeier, R. and Tribble, G., 1988. Well point sampling of reef interstitial water. Coral Reefs, 7: 19–22.
Smith, S.V. and Buddemeier R.W., 1992. Global change and coral reef ecosystems. Annu. Rev. Ecol. Syst., 23: 89–118.
Tribble, G., Sansone, F., Smith, S., 1990. Stoichiometric modeling of carbon diagenesis within a coral reef framework. Geochim. Cosmochim. Acta, 54: 2439–2449.
Trichet, J. and Fikri, A., 1993. Information given by organic matter on the origin of insular phosphorites. Inter. Symposium on Phosphogenesis. Interlaken (abstract).
Underwood, M.R., Peterson, F.L. and Voss, C.I., 1992. Groundwater lens dynamics of atoll islands. Wat. Resour. Res., 28 (11): 2889–2902.
Wauthy, B., 1986. Physical ocean environment in the South Pacific Commission Area. UNEP Reg. Seas Reports and Studies, 83, 90 pp.
Whitaker, F. and Smart, P., 1990. Active circulation of saline ground waters in carbonate platforms: evidence from the Geat Bahama Bank. Geology, 18: 200–203.
Whitehead, N.E., 1993. The elemental content of Niue island soils as an indicator of their origin. N.Z. J. Geol. Geophys., 36: 243–254.
Wolanski, E., Drew, E., Abel, K. and O'Brien, J., 1988. Tidal jets, nutrient upwelling and their influence on the productivity of the alga Halimeda in the ribbon reefs. G.B.R. Estuar. Coast. Shelf. Sci., 26: 169–201.

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