DIAGENESIS OF NEOGENE DELTA-FRONT PATCH REEFS: ALTERATION OF COASTAL, SILICICLASTIC-INFLUENCED CARBONATES FROM HUMID EQUATORIAL REGIONS

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ABSTRACT: This study offers insights into the diagenetic alteration of coastal carbonates that formed coevally with nearly continuous siliciclastic influx in a humid equatorial setting. A multi-disciplinary petrographic, cathodoluminescent, stable-isotope, trace-element and major-element investigation allowed characterization of diagenetic features, paragenetic sequencing, and an interpretation of diagenetic environments from Neogene patch reefs of the Samarinda region, Mahakam Delta, Borneo, SE Asia. Marine cements are absent from the patch reefs, with grain micritization the only marine diagenetic feature recognized. The predominant diagenetic feature within the patch reefs is pervasive neomorphic stabilization and cementation of aragonite reef components to calcite that pre-dates all compaction features. Meteoric aquifer flow derived from the adjacent landmass is inferred as the main parent diagenetic fluid, since δ18O V-PDB values of calcite cements of −3.6 to −11.7‰ are consistent with precipitation from SE Asian freshwater, and inconsistent with a wholly marine origin. Late-stage fracturing, cementation, and chemical compaction are relatively minor features and attest to a changing paleohydrologic and diagenetic environment. Evidence for a shallow to moderate burial diagenetic regime for these later features are maximum temperature of 53°C and burial depths < 1000 m inferred from stable-isotope values of calcite and late dolomite cements, dolomite crystal fabrics, the onset depth of styololite or dissolution-seam formation, and regional geothermal gradients. The humid tropical environment and “ever-wet” conditions on the island of Borneo together with rapid Cenozoic uplift likely led to paleoaquifer flow with fluids focused through adjacent deltaic units into the reef carbonates. In these coastal carbonates from the humid Samarinda region, continental groundwater flow has resulted in pervasive stabilization and calcitization, features rare in arid or temperate counterparts.

INTRODUCTION

Carbonate–siliciclastic mixing is common in both modern and ancient environments (Mount 1984; Larcombe and Woolfe 1999; Woolfe and Larcombe 1999; Wilson and Lokier 2002; Wilson 2005). In spite of their common occurrence, the diagenesis of siliciclastic-influenced coral reefs and carbonate platforms from equatorial to subtropical settings remains largely unstudied (Hendry et al. 1999; Wilson 2012). The hypotheses tested here are that: (1) the diagenesis of continental margin reef deposits in a region of siliciclastic input will be influenced strongly by basin-margin paleohydrology, and (2) the alteration of coastal carbonates from other climatic regions (cf. Dodd and Nelson 1998; Hendry et al. 1999; Ricketts et al. 2004; Caron and Nelson 2009) may not be directly analogous to that of comparable deposits in the equatorial tropics.

To test these hypotheses, this study evaluates the diagenesis of Miocene delta-front reefs from Borneo, with three aims: (1) to characterize the extent of diagenesis in these mixed carbonate–siliciclastic patch reefs from a coastal setting, (2) to develop a paragenetic sequence of the succession in the context of the evolving basin hydrology and the climatic setting, and (3) to compare general aspects of diagenesis of these humid equatorial carbonates with that of other coastal carbonates from humid-nontropical and arid settings. Here, “coastal carbonate” refers to nearshore marine carbonate development along predominantly siliciclastic coastlines, such as in delta-front, fan-delta, or siliciclastic inner-shelf settings. The results of this study provide insights into patterns of diagenesis in coastal carbonate systems that may have broad analogs elsewhere (Hendry et al. 1999; Nelson et al. 2003; Wilson and Hall 2010; Wilson 2012).

Neogene carbonates have developed extensively in the tropical marine waters surrounding Borneo, despite abundant sediment runoff from Borneo, the largest island in Indonesia (Fig. 1; Wilson et al. 1999; Hall 2002a). Modern and Neogene carbonates exist coevally with major siliciclastic influx from the Mahakam Delta on the eastern side of Borneo (Fig. 1; Roberts and Sydow 1996; Wilson and Lokier 2002; Hook and Wilson 2003; Wilson 2005). Outcrops, with up to 90% exposure, of Miocene patch reefs interbedded with Mahakam Delta deposits crop out onshore East Borneo (Fig. 2; Alam et al. 1999; Wilson 2005). The sedimentology and biostratigraphy of these units is well documented, as is the siliciclastic influence on carbonate production, reef growth, and sequence development (BouDagher-Fadel and Wilson 2000; Wilson and Lokier 2002; Wilson 2005; Lokier et al. 2009). The present study extends these results, focusing on the diagenesis of these deposits, by combining outcrop data with petrography, cathodoluminescence, and geochemistry.

GEOLOGICAL SETTING

Tectonic plate convergence and weathering under humid equatorial climatic conditions in Southeast Asia have resulted in high rates of uplift and erosion of the Borneo landmass (Hall 1996; Hall and Nichols 2002).
Since at least the Neogene, large volumes of siliciclastic sediments derived from Borneo have been deposited into the deep oceanic basins to the north and east of the island, with associated rapid delta progradation (Fig. 1; Hamilton 1979; Wilson and Moss 1999; Hall and Nichols 2002). The Mahakam Delta has been contributing to the Cenozoic sedimentation of the Kutai Basin, in Eastern Borneo, since at least the early Miocene (van de Weerd and Armin 1992; Allen and Chambers 1998; Moss and Chambers 1999). The eastern Kutai Basin has a Neogene sedimentary section of at least 9 km thickness (Hamilton 1979; Cloke et al. 1999; Hall and Nichols 2002).

The Kutai Basin is one of a series of sedimentary basins that developed around the margins of the Eurasian continental crust during the Paleogene (van de Weerd and Armin 1992; Hall 1996). The cause and exact timing of regional basin initiation is still debated, but most authors have inferred that the Kutai Basin has rift-related extensional or transtensional origins (van de Weerd and Armin 1992; Moss et al. 1997; Moss and Chambers 1999). Covering a region approximately 165,000 km², the Kutai Basin (Fig. 1) is the largest and deepest basin in Indonesia, separated from the Tarakan Basin in the north and the Barito Basin in the south by the WNW–ESE-trending Mangkalihat Ridge and...
Adang Flexure, respectively (van de Weerd and Armin 1992; Alam et al. 1999; Moss and Chambers 1999). Throughout the Cenozoic, carbonates developed on faulted highs at the northern (Fig. 1; e.g., the Taballar Formation) and southern (e.g., the Berai Limestone) margins of the Kutai Basin, as well as faulted highs largely isolated from siliciclastic influx within the basin (e.g., the Kidango Limestone and the Ritan Limestone; Moss and Chambers 1999; Wilson et al. 1999). Subsequent to regional subsidence, several Neogene inversion episodes affected the basin, resulting in pulses of rapid delta progradation (Allen and Chambers 1998). Despite a high siliciclastic flux into the Mahakam Delta, Neogene to modern carbonates formed as proximal delta-front patch reefs and more distal shelf accumulations (Roberts and Sydow 1996; Wilson 2005; Saller et al. 2010). These Neogene delta-front patch reefs are the subject of this study.

Based on foraminifera and nannofossils, the western delta-front patch reefs at Senoni and Kota Bangun are of early Miocene age, older than the early-middle Miocene deposits cropping out to the east at Airputih, Dibelakan Parliament (DPR), Permasip, Bontang, and Badak (Fig. 3; BouDagher-Fadel and Wilson 2000; Wilson 2005). These Miocene reef deposits are analogous in size, location, and lithology to delta-front proximal patch reefs, and to a lesser extent more distal shelf carbonates, from the modern and Pleistocene of the Mahakam Delta. These post-Neogene carbonates create platforms and mounds on the northern part of the shelf (Roberts and Sydow 1996). Bioherms in the inner and middle shelf are up to 40 m thick, but they are currently inactive and are covered in a meter-scale layer of clay (Roberts and Sydow 1996). The late Pleistocene-to-modern development of carbonates has been controlled by two factors: (1) the northern delta lobe has a lower rate of sediment supply than the southern lobe, and (2) the Indonesian Throughflow Current flows south at high speed. Both factors favor waters less turbid in the north than in the south (Roberts and Sydow 1996). Miocene carbonates also were concentrated towards the north, and (2) the Indonesian Throughflow Current flows south at high speed. Both factors favor waters less turbid in the north than in the south (Roberts and Sydow 1996).

**MORPHOLOGY AND FACIES OF THE MICHOEAN PATCH REEFS**

The outcrops of Miocene patch reefs studied here are up to 2–4 km wide with a post-compactional thickness of up to 40 m (Fig. 3; Wilson 2005). The patch reefs overlie, interfinger with, and are overlain by siliciclastic deposits of the Mahakam Delta (Wilson 2005). Corals, many in growth position, individually surrounded by a clay-rich and micritic groundmass, dominate the patch-reef deposits. Bioclastic grains including larger benthic foraminifera, coralline algae, mollusks, echinoderms, and *Halimeda* locally contribute to the deposits. None of the samples contain less than 5 weight % admixed siliciclastics, although they may be as high as 80%, suggesting siliciclastic influx contemporaneous with patch-reef development (Wilson 2005). Distinctive zoning of biota and facies within individual patch reefs suggests a dynamic interrelationship between siliciclastic input, nutrient influx, and the carbonate producers (Wilson and Lokier 2002; Wilson 2005). Twelve facies were defined from outcrop and petrographic studies on the basis of their constituent components and their textural and lithological features (Wilson 2005). These facies have been grouped into three facies associations, on the basis of their sedimentary and biotic characteristics: (1) siliciclastic facies (> 95% siliciclastics), (2) mixed carbonate-siliciclastic facies (35–80% siliciclastics), and (3) carbonate facies (6–35% siliciclastics; Wilson 2005). The siliciclastic facies association is restricted to the deltaic system and is not considered in the present study. Within individual patch reefs a trend of decreasing siliciclastic content is associated with a change from platy to branching to massive corals towards the core of the patch reefs (Figs. 3, 4). Specifically, mixed carbonate-siliciclastic facies (including common platy coral sheetstones; nomenclature of Insalaco 1998) dominate at the base, top, and margins of the patch reefs and bordering areas where they interfinger with siliciclastics of the delta. Towards the center of the patch reef, carbonate facies change successively from branching coral pillarstones and coral rud-floatstones to head coral mix-domestones as siliciclastic content decreases. Deposits richer in disseminated carbonaceous plant remains include more abundant calcareous algae, echinoderms, and locally larger benthic foraminifera (Wilson and Lokier 2002; Wilson 2005). The change in the faunal assemblage is consistent with a dominance of calcareous algal communities over framework-building reef communities associated with nutrient loading in the modern and the Quaternary (Roberts et al. 1988; Roberts and Sydow 1996). Siliciclastics within the patch-reef deposits are predominantly clay (including kaolinite and smectite) and subsidiary fine sand (Wilson 2005).

**METHODS**

Mixed carbonate-siliciclastic deposits are exposed in active and disused quarries and unquarried karst through seven patch-reef complexes in
Eastern Borneo (Figs. 2, 3; Wilson 2005). Up to 90% exposure allowed high-resolution sampling, outcrop logging, facies mapping, and section correlation through individual patch reefs (Wilson 2005). Twenty-three sections through the seven reef complexes were logged, and of the 250 samples collected, 117 have been studied petrographically (Appendix 1, see Acknowledgments section).

Lithological components, microfacies, diagenetic phases, and the relative timing of diagenetic events were determined through petrographic analysis of the thin sections. Half of each thin section was stained with Alizarin Red S and potassium ferricyanide for identification of dolomites, ferroan calcite, and nonferroan calcite in thin section. The relative abundance of components and diagenetic phases were recorded semi-quantitatively (visual estimates: Appendix 1), with the results closely comparable with fully quantitative point counting and acid digestion analyses previously undertaken on 52 of the samples (Wilson 2005; Lokier et al. 2009). Cold cathodoluminescent (CL) microscopy study of 26 polished thin sections utilized a Technosyn 8200 MKII luminoscope (after Witkowski et al. 2000). Samples for CL analysis, selected primarily from thin sections with a variety of coarse (0.25 mm) cements, facilitated investigation of the range of cement phases present. On the basis of the relative timing of petrographically defined diagenetic features, three groups are evident (Appendix 1): (1) pre-fracture, (2) fracture-associated, and (3) post-fracture features.

Stable-isotope analysis ($\delta^{18}O$ and $\delta^{13}C$) is based on 38 samples micro-drilled from the rock off-cut counterpart of thin sections. Drilled samples included ferroan and nonferroan cements, and fracture (~1 mm) filling cements, dolomites, and bioclastic components. Twenty oxygen and carbon isotope analyses were run on a VG Isocarb automated system online to a VG Isogas Prism II isotope-ratio mass spectrometer at SUERC, East Kilbride. All data have been normalized to NBS-19, and are reported relative to V-PDB. Replicated analyses of an internal carbonate standard (Mab2b) were reproducible to ±0.1‰. An additional 18 oxygen and carbon isotope analyses (samples with the prefix Madden) were run on a GasBench II system coupled online to a stable-isotope-ratio mass spectrometer in continuous flow (Skrzypek and Paul 2006), with all data normalized to NBS-19 and reported relative to V-PDB; these analyses were undertaken at UWA, Perth. External errors for $\delta^{18}O$ and $\delta^{13}C$ were 0.20‰ and 0.10‰, respectively.

Trace-element and major-element analyses on five cement samples and five matrix samples micro-drilled from hand specimen were subject to a four acid (hydrofluoric, nitric, perchloric, and hydrochloric) digestion process. Hydrofluoric acid was used to allow comparison of elemental signatures in the pure carbonate cements (containing no clay) and those of the matrix with admixed micrite and clays. A correction was not applied to account for the dissolution of the silicate fraction. Trace-element and major-element signatures were detected by inductively coupled plasma optical emission spectrometry (ICP-OES). The results were within normal limits (WPR-1 reference standard) and no normalization was required. The results are reported as molar concentrations of the original sample. Precision for each sample is reported as

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**FIG. 3.—** Simplified sedimentary logs of patch reefs from the Samarinda region in east Borneo showing their dimensions, ages, and characteristics. The lateral distances between patch reefs are not to scale (from Wilson 2005).
FIG. 4.—Measured sections through the Airputih patch reef showing key lithologies and facies associations (from Wilson 2005) with petrographically defined diagenetic attributes and cathodoluminescent characteristics. On each measured section, to the right of the lithology column, the diagenetic features in individual thin sections are shown in order of occurrence (left to right—oldest to youngest), with their CL characteristics, where studied, shown directly above. Single letters indicate one CL character, two letters separated by "&" indicates two common characteristics, and two letters separated by "-" indicates a transition from one character to the other (modified from Wilson 2005).
percentage variability at the 95% confidence limit for each element. Precision values have a low variation amongst samples; these values range from 4.5 to 9.3% for Al, 4.2% for Ca, 4.3 to 6.5% for Fe, 4.3 to 6.4% for K, 4.3 to 5.4% for Mg, 4.9 to 6.2% for Mn, 4.4 to 5.4% for Na, and 4.7% for Sr. Variance is greater for S, with a range of 5.8 to 29.5%.

DIAGENETIC CHARACTERISTICS OF THE PATCH REEFS

Petrography of Pre-Fracture Features

Micritization.—Micritization of bioclastic components is prevalent in 114 of 117 thin sections. Micrite envelopes are visible in plane-polarized light (PL) as narrow (10-60 μm) dark brown rims to bioclasts. Corals show the most pervasive micritization, with rims up to 60 μm thick, and *Halimeda* showing rims of only 10 μm width. The CL properties of the originally calcitic bioclasts (e.g., coralline algae and larger benthic foraminifera) are typically bright to dull luminescent. This CL signature contrasts with that of the micrite envelopes, which are very dull to nonluminescent. The CL character of micritized features is consistent between, and within, individual patch reefs.

Cone-in-Cone Calcite Cementation.—Cone-in-cone calcite, present in approximately 10% of thin sections, is typically nonferroan and is rarely (~ 5% of samples) a dominant cement phase. Cone-in-cone calcite post-dates micritization, but it formed prior to the granular mosaic calcite (C1), since the latter replaces portions of the cones (Fig. 5A). This cement is present in samples with abundant fine siliclast-rich micritic matrix (25-50% of total rock volume). Cone-in-cone cement has developed within the matrix, and commonly close to the periphery of bioclasts, most notably corals. At Airputih and Permasip (Figs. 3, 4), cone-in-cone cement is confined to samples from either the top or basal portions of the patch reefs. In comparison, at the < 20-m-thick DPR section, this cement is pervasive throughout. The cone-in-cone cement has a divergent fibrous structure, with cone-shaped bundles of crystalline calcite columns and a displacive habit. Individual crystals are variable in size, typically ranging from 10 to 100 μm in width and up to 500 μm in length. Crystals have an even and regular structure with regular crystal boundaries. Individual crystals are colorless to pale yellow to brown in plane light and are separated by a < 10-μm-thick coating of clay minerals or micrite. Limited CL imagery reveals that in at least one sample, cone-in-cone cements have the same dull to moderate luminescence as the granular mosaic cements, but with a < 10-μm-thick, brightly luminescent rim to bundles of cones.

Granular Mosaic and Blocky Calcite (C1).—Mosaic and blocky calcite cement is a dominant diagenetic feature (up to 50% of the thin section), present in 98% of thin sections. This cement is confined to bioclastic components showing relics of the original skeletal structure or occurs as minor mosaic cement extending into the chambers of bioclasts (Fig. 5B). The cement crystals show a wide variability in size, and where they replace original coral fabric or structure crystal sizes range from 100 to 200 μm up to 600 to 700 μm. The variable but commonly coarse texture contrasts with crystal mosaic found in the chambers of other bioclasts, such as larger benthic foraminifera, where crystal sizes are mostly 100–150 μm. Localized areas of mosaic to granular cement are typified by a uniform mosaic of crystals with irregular crystal boundaries, and little or no evidence of drusy cement morphology. Relicts of the internal skeletal structures and walls of corals are partially preserved through original organic or inorganic inclusions and outlining micrite envelopes. The majority of mosaic cements are nonferroan (stained pink with Alizarin Red S and potassium ferricyanide), with a rare, slight ferroan signature (pale blue stain) in ~ 15% of thin sections. In plane light, these cements are colorless to gray, and have moderate-dull luminescence to non-luminescence in CL or bright CL where C1 blocky calcite grows into primary pore spaces. CL also reveals that these crystals grow crosscutting the micrite envelope and into areas of primary intraparticle porosity within bioclast chambers as outlined by the micritic envelope (Fig. 6A/A*).

Petrography of Fracture-Associated Features

Grain Breaks and Fractures (F1).—Breakage is accommodated via minor crystal or grain fractures, typically on the order of hundreds of μm in length and no more than 20 μm in width. There is minor to no offset of fractured components. The full extent of minor early fracturing is most evident in CL as fine fracture networks filled with dull-luminescent to nonluminescent C2 cements that crosscut earlier dull-luminescent C1 cement, and bright-luminescent primary depositional features including micritic matrix (Fig. 6A/A*).

Blocky to Equant Calcite (C2 and C3).—Thirty-five percent of thin sections show C2 calcite, of which 76% are weakly ferroan to ferroan and 24% are nonferroan. C2 cement fills F1 fractures with some associated growth into, and incorporation of, areas of micritic matrix (Fig. 5C), whereas C3 cements fill later F2 and F3 fractures (Fig. 5D). Eighty-three percent of thin sections show a C3 calcite, of which 40% are weakly ferroan to ferroan and 60% are nonferroan. In plane light, C2 and C3 cements are colorless to pale yellow and are dull luminescent to nonluminescent in CL. However, C2 cement may show zoning from bright to dull luminescence in larger pores (Fig. 6B/B*). Both C2 and C3 cements have crystal sizes ranging from 100 to 600 μm, lack a preferred orientation, and have regular to subcircular crystal boundaries.

Grain Compaction and Fractures (F2, F3, and F4).—Following C2 cementation, a nonpervasive episode of grain compaction and fracturing (F2) occurs, and is evident in ~ 55% of thin sections. F2 fractures have thicknesses of 10 to 100 μm and crosscut matrix (commonly close to bioclasts), bioclasts, and earlier cements. A third stage of grain fracturing (F3) is recognized in 60% of thin sections, where F3 fractures crosscut F2 fractures (Fig. 5D). This third set of fractures is more numerous and may form “thread-works” of fractures in areas of micritic or siliclastic matrix (Fig. 5D). Both F2 and F3 fractures are filled by C3 cement with dull luminescence to nonluminescence in CL. Compaction and fracturing did not lead to alignment, rotation, or flattening of grains. F4 fracturing is rare (10% of thin sections), nonextensive and nonpervasive, with fractures typically extending < 500 μm in length and a few tens of μm in width. F4 fractures crosscut earlier cements or, more commonly, contribute to the breakage of previously cemented fractures. F4 fractures are not cemented and remain open.

Petrography of Post-Fracture Features

Stylolites and Dissolution Seams.—Stylolites and dissolution seams are found in 23% and 88% of thin sections, respectively. Stylolites appear as serrated seams (Fig. 5D), and dissolution seams are smooth and undulating (Fig. 5E). Stylolites have amplitudes of up to 1 mm, are several centimeters in length, and are found in areas with minor siliclastic material in the matrix and with greater cementation. Stylolites crosscut bioclasts, earlier cement phases, and pre-existing fractures. Dissolution seams occur in samples with greater micritic or siliclastic-rich matrix. Individual samples may have stylolites in calcite-cemented regions that change along length to dissolution seams in matrix-rich areas. Dissolution seams also crosscut older cement phases and fractures, but typically they do not intersect bioclasts, instead commonly going around or between them (Fig. 5E). Both features are identified through the concentration of insoluble noncarbonate material, resulting in a dark brown seam in plane light (Figs. 5D, E).
FIG. 5.—Plane-polarized light (PL) thin-sections photomicrographs illustrating a range of diagenetic features in patch-reef deposits. A) Cone-in-cone calcite cement showing fibrous structure arranged into bundles. Darker seams are concentrations of micritic–siliciclastic material. Granular mosaic C1 replacement of cones occurs at the top of the field of view, with the white arrow showing a replacement front. B) Granular mosaic C1 calcite replacing coral chamber walls (1), and contemporaneous blocky calcite cementation of primary porosity (2). C) F1 fractures crosscutting micritic and clay-rich matrix and probable serpulid worm chamber infill. Fractures are filled by a later stage of C2 calcite cement. Blocky to equant C2 cement includes localized inclusions of dark-colored micritic matrix, attributed to aggrading neomorphism of the matrix. D) Two stages of fractures F2 (arrow 1) and F3 (arrow 2) crosscutting mixed micritic–siliciclastic matrix and filled with C3 calcite cement. Fractures are crosscut by a late-stage stylolite (arrow 3). E) Dissolution seam (arrows) crosscutting micritic matrix and encrusting coralline algae between two corals. F) Micro-dolomite rhombs partially replacing the micritic chamber fill of a coral that has undergone granular mosaic C1 replacement.
FIG. 6.—PL and CL photomicrograph pairs illustrating different diagenetic features. A/A*) Brightly luminescent micrite fill of coral chamber (1), dull-luminescent granular mosaic C1 replacement of a coral wall (2), contemporaneous blocky C1 calcite growth into cavity with nonluminescence (3) and dull-luminescent C2 calcite filling pervasive fracturing not evident in PL (4). B/B*) Pervasive micritization replacing an originally brightly luminescent bioclast (1), dull-luminescent character of micritic sediment (2) and zoning of C2 blocky calcite cement from bright luminescence (3) into nonluminescence at the center of the cement fill (4). C/C*) Detailed changes within a coral, with bright CL character of micritic chamber infill (1), C1 granular neomorphic cement with dull luminescence to nonluminescence (2) and C1 blocky cement fill of coral chamber (3) showing weak zoning with bright luminescence at the edge of some crystals towards a bright-dull center.
Dolomite Cement.—Dolomite rhombs are found in < 15% of thin sections, but where present, they commonly are abundant. Mosaics of euhedral to subhedral 10–60 µm micro-dolomite rhombs with regular crystal boundaries typically occur within the micritic matrix infilling bioclast chambers (Fig. 5F). Rhombs are also present along dissolution seams and are distributed in the matrix close to seams. In plane light, dolomites are colorless to pale yellow or brown, and in few samples have a cloudy appearance. In CL, they appear as brightly luminescent “spots” or, rarely, dull-luminescent rhombs. There are no crosscutting relationships that suggest that the dolomites were affected by later diagenetic phases.

Cathodoluminescence

CL characteristics are variable at scales from across individual patch reefs (Fig. 4) to within isolated corals (Figs. 6A/A*, C/C*), but some trends are evident. For example, the Airputih and Permasip patch reefs show consistent trends, including: micritic areas typified by bright to dull luminescence, C1 cement with a dull luminescence, and C2/C3 cements with dull luminescence to nonluminescence. There is also a slight tendency for the CL character of carbonate facies (< 35% siliciclastics) to have a brighter luminescence in general than that of mixed carbonate–siliciclastic facies (35-80% siliciclastics). Consistent CL trends are not evident in the patch reefs of Senoni, Bontang, Badak, or Dibelekan Parliament, although CL study was less extensive.

Matrix or cements within large coral chambers have CL characteristics generally brighter than the reef edge or reef core and may show weak zoning within these coral “micro-environments.” This trend is evident in C1 cements where nonluminescent C1 cement replacing the coral skeleton changes into dull-luminescent to bright-luminescent pore-filling micrite and blocky C1 cement (Fig. 6C/C*).

Stable Isotopes

Stable-isotope composition (Table 1, Fig. 7) of 31 calcite and six dolomite samples define a tight group, with low variance among individual values of δ18O. These values, ranging from −3.6‰ to −11.7‰ V-PDB, are similar to more negative than that expected from cements derived from Oligo-Miocene southeast Asian marine waters (δ18O values of −1.5 to −4.2‰; Ali 1995; Wilson and Evans 2002). δ13C values also lie within a narrow field, generally within ± slightly more negative than, the expected range of normal marine waters (+1‰ to −1‰ δ13C V-PDB; Ali 1995; Wilson and Evans 2002), where values range from +1.1‰ to −2.6‰ δ13C V-PDB for all but two samples. Both of the two outlying values at −3.6‰ and −6.2‰ δ13C V-PDB are from C2 blocky cements.

Trace Elements and Major Elements

The element ratios of magnesium to calcium (Table 2) indicate mol % close to a value of 4 mol % Mg, with two C1 and two micritic samples showing a slightly higher concentration of 6 to 7 mol % Mg (intermediate calcite; James 1997). A correction was not applied to account for the dissolution of the silicate fraction in the matrix; therefore Mg contents may reflect both the carbonate and siliciclastic content of the matrix. Data reveal several distinctive trends for the trace-element and major-element results within the paired samples of C1 calcite and their associated micritic-siliciclastic matrix. Mn/Ca and Sr/Ca ratios show little variability from matrix to C1 calcite. In contrast, the ratios of Na/Ca, Fe/Ca, and Mg/Ca decrease from the matrix to C1 samples. Aluminum content is highly variable, with values of ~1 to 13 mol % Al in C1 samples and ~27 to 72 mol % Al in micrite. Ratios of Al/Ca are consistent, showing a decreasing concentration from matrix to C1 (Table 2).

Diagenetic, Temperature, and Paleohydrology Interpretations

Collectively, the observations reveal a paragenetic sequence with several phases of cementation within the Samarinda patch reefs (Fig. 8). The Anderson and Arthur (1983) equation (Equation 1) provides a means to derive Miocene δ18O seawater values for the region:

\[
T = 16 - 4.14(\delta^{18}O_{\text{Calcite}} - \delta^{18}O_{\text{Seawater}}) + 0.13(\delta^{18}O_{\text{Calcite}} - \delta^{18}O_{\text{Seawater}})^2
\]

A Miocene δ18O seawater value for the region of −2 to 0‰ V-SMOW has been derived using this equation and the observed range of calcitic bioclast values for SE Asian Oligocene-Miocene seawater and an assumed ocean surface temperature of 25 °C (Neogene of coastal Borneo; Ali 1995). The larger benthic foraminifera in the Samarinda samples (δ18O V-PDB values of −5.8‰ to −10.0‰) have lower δ18O values than the known range for SE Asian Oligocene-Miocene calcitic bioclasts, perhaps reflecting: (1) the samples containing contaminating matrix, (2) precipitation out of equilibrium with Miocene sea water, or (3) resetting of the δ18O during stabilization by meteoric waters or elevated temperatures. Because the deposition of Neogene patch reefs was impacted by terrestrial runoff, the strongly negative δ18O V-PDB values of the larger foraminifera may partially reflect a brackish signature. Given the uncertainties of the isotopic signatures of the larger foraminifera in the Samarinda patch reefs reflecting marine signatures, estimated Miocene seawater values are used (−1.5‰ to −4.2‰ δ18O V-PDB; Ali 1995; Wilson and Evans 2002) in deriving a regional V-SMOW seawater value. A δ18O value of −4 to −8‰ V-SMOW is suggested for meteoric parent fluids on the basis of δ18O values of meteoric precipitation in SE Asia of −4 to −6‰ at low elevations (Bowen and Wilkinson 2002) and up to −8‰ for the whole of Borneo (Anderson and Arthur 1983).

The onset of development of styloites and dissolution seams occurs in moderate-burial to deep-burial environments of 500–1000 m (Nicolaides and Wallace 1997; Machel 2004). Since all calcite cement precipitation pre-dates development of styloites and dissolution seams, a maximum burial depth for calcite cementation of the Samarinda patch reefs of 1000 m is inferred. Extrapolation of the eastern Kutai Basins geothermal gradient (27.5 °C/km; Hall 2002b) gives an estimated maximum calcite precipitation temperature of 53 °C.

Interpretation of Pre-Fracture Features

Micritization.—Micritization present throughout the samples is the first diagenetic process, on the basis of crosscutting relationships by each of the later features. Skeletal material in shallow-marine environments commonly is micritized due to micrite infilling the micro-boring of endolithic organisms (Bathurst 1966; Gunther 1990; Perry 1999). In particular, in warm and low-energy environments, as is inferred for these samples (Wilson 2005), micritization is commonly recognized as an early marine diagenetic feature (Gunther 1990; Perry 1999). The very dull-luminescent to nonluminescent CL character of the micritic envelopes is consistent with associated marine or oxidizing pore fluids. Alternatively the fine texture of the micritic material may impart a darker CL character.

Cone-in-Cone Calcite.—Cone-in-cone calcite cement formed prior to the onset of C1 cementation. Well-formed bundles of crystals within the micritic-siliciclastic substrate indicate that the cone-in-cone cement was precipitated prior to lithification of the matrix (cf. Franks 1969; McBride et al. 2003). The predominant distribution of this cement at the top and base of patch reefs adjacent to siliciclastics and in siliciclastic-rich units (Fig. 4) may reflect enhanced fluid flow and diffusion from the more...
permeable siliciclastics, promoting precipitation of the cement (cf. McBride et al. 2003). A dull to moderate CL character of the cone-in-cone cement is suggestive of oxidizing to perhaps slightly reducing conditions (iron derivation from the siliciclastics may have contributed to reducing conditions). δ18O V-PDB values of −8.5 to −10.2‰ for the cone-in-cone calcite, if precipitated from marine parent fluids, would require temperatures of 50–70 °C at a burial depth of approximately 500–1600 m. This temperature and depth range is unlikely because the cone-in-cone cement precipitated prior to mechanical compaction features (below), and under conditions that were still oxidizing or perhaps slightly reducing. At temperatures of 25 to 35 °C, consistent with shallow burial, the cone-in-cone cements would have formed from pore fluids with δ18O values of −7.3 to −5.3‰ V-SMOW (Fig. 9), suggestive of meteoric fluids with stable-isotope values ranging from −6.4 to −11.7‰ δ18O V-PDB, would require temperatures of 40–80 °C, at a burial depth of approximately 500–1700 m. These temperatures and depths are highly unlikely due to the pre-fracture timing of the C1 cement and potentially still oxidizing precipitation conditions. A formation fluid of −7 to −5‰ V-SMOW at 25 to 35 °C is likely (Fig. 9), consistent with early meteoric to shallow-burial diagenesis and similar to the conditions inferred for the cone-in-cone cement that this C1 cement partially replaces. However, samples with low δ18O values may be indicative of: (1) mixing marine pore fluids with meteoric water, or (2) meteoric diagenesis at higher rock:water ratios. With high δ18O values may be indicative of: (1) mixing marine pore fluids with meteoric water, or (2) meteoric diagenesis at higher rock:water ratios. With high δ18O values may be indicative of: (1) mixing marine pore fluids with meteoric water, or (2) meteoric diagenesis at higher rock:water ratios. With high δ18O values may be indicative of: (1) mixing marine pore fluids with meteoric water, or (2) meteoric diagenesis at higher rock:water ratios. With high δ18O values may be indicative of: (1) mixing marine pore fluids with meteoric water, or (2) meteoric diagenesis at higher rock:water ratios. With high δ18O values may be indicative of: (1) mixing marine pore fluids with meteoric water, or (2) meteoric diagenesis at higher rock:water ratios. With high δ18O values may be indicative of: (1) mixing marine pore fluids with meteoric water, or (2) meteoric diagenesis at higher rock:water ratios.

**Table 1.** δ13C‰ V-PDB and δ18O‰ V-PDB values from samples of the Samarinda patch reefs.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Name</th>
<th>δ13C‰ V-PDB</th>
<th>δ18O‰ V-PDB</th>
<th>Phase</th>
<th>Comment</th>
</tr>
</thead>
<tbody>
<tr>
<td>AB01</td>
<td>Madden_AB01</td>
<td>−0.74</td>
<td>−10.01</td>
<td>Blocky equant cement, continuation of granular mosaic (C1)</td>
<td></td>
</tr>
<tr>
<td>AA01</td>
<td>AA01*1.0</td>
<td>−1.43</td>
<td>−9.46</td>
<td>Granular mosaic cement (C1)</td>
<td></td>
</tr>
<tr>
<td>AE05a</td>
<td>AE05a*2.1</td>
<td>0.91</td>
<td>−7.20</td>
<td>Granular mosaic cement (C1)</td>
<td></td>
</tr>
<tr>
<td>AC08</td>
<td>Madden_AC08</td>
<td>−0.34</td>
<td>−9.15</td>
<td>Cone-in-cone cement</td>
<td></td>
</tr>
<tr>
<td>AB12a</td>
<td>AB12a*1.0</td>
<td>−1.79</td>
<td>−10.41</td>
<td>Granular mosaic cement (C1)</td>
<td></td>
</tr>
<tr>
<td>AB07</td>
<td>AB07*1.3</td>
<td>−6.22</td>
<td>−3.61</td>
<td>Blocky equant cement (C2)</td>
<td>slightly ferroan</td>
</tr>
<tr>
<td>AA04a</td>
<td>AA04a*1.1</td>
<td>−1.60</td>
<td>−6.43</td>
<td>Granular mosaic cement (C1)</td>
<td></td>
</tr>
<tr>
<td>SB10</td>
<td>Madden_SB10</td>
<td>−0.96</td>
<td>−6.87</td>
<td>Larger benthic foraminifera</td>
<td></td>
</tr>
<tr>
<td>AE02</td>
<td>Madden_AE02</td>
<td>−0.09</td>
<td>−9.96</td>
<td>Larger benthic foraminifera</td>
<td></td>
</tr>
<tr>
<td>AB08</td>
<td>Madden_AB08</td>
<td>0.42</td>
<td>−5.79</td>
<td>Larger benthic foraminifera</td>
<td></td>
</tr>
<tr>
<td>AK12</td>
<td>Madden_AK12</td>
<td>1.11</td>
<td>−8.17</td>
<td>Larger benthic foraminifera</td>
<td></td>
</tr>
<tr>
<td>AA26</td>
<td>AA26*0.9</td>
<td>0.57</td>
<td>−7.86</td>
<td>Larger benthic foraminifera</td>
<td></td>
</tr>
<tr>
<td>AC08</td>
<td>Madden_AC08</td>
<td>−0.34</td>
<td>−9.15</td>
<td>Cone-in-cone cement</td>
<td></td>
</tr>
<tr>
<td>AL02</td>
<td>Madden_AL02</td>
<td>−0.23</td>
<td>−10.21</td>
<td>Cone-in-cone cement</td>
<td></td>
</tr>
<tr>
<td>AA20</td>
<td>AA20*0.8</td>
<td>−0.79</td>
<td>−7.98</td>
<td>Granular mosaic cement (C1)</td>
<td></td>
</tr>
<tr>
<td>AB04</td>
<td>Madden_AB04</td>
<td>−1.82</td>
<td>−10.74</td>
<td>Vein (Blocky C3 cement filling fracture)</td>
<td></td>
</tr>
<tr>
<td>AB06</td>
<td>Madden_AB06</td>
<td>−0.76</td>
<td>−10.49</td>
<td>Vein (Blocky C3 cement filling fracture)</td>
<td></td>
</tr>
<tr>
<td>BB05</td>
<td>BB05*0.5</td>
<td>−3.60</td>
<td>−4.86</td>
<td>Blocky equant cement (C2)</td>
<td></td>
</tr>
<tr>
<td>AB12a</td>
<td>AB12a*1.0</td>
<td>−1.52</td>
<td>−10.08</td>
<td>Granular mosaic cement (C1)</td>
<td></td>
</tr>
<tr>
<td>AA15</td>
<td>Madden_AA15</td>
<td>−0.76</td>
<td>−10.49</td>
<td>Vein (Blocky C3 cement filling fracture)</td>
<td></td>
</tr>
<tr>
<td>PC03</td>
<td>Madden_PC03</td>
<td>−0.41</td>
<td>−7.30</td>
<td>Vein (Blocky C3 cement filling fracture)</td>
<td></td>
</tr>
<tr>
<td>KB07</td>
<td>KB07*0.9</td>
<td>−0.15</td>
<td>−10.04</td>
<td>Vein (Blocky C3 cement filling fracture)</td>
<td></td>
</tr>
<tr>
<td>AA04</td>
<td>Madden_AA04</td>
<td>−2.62</td>
<td>−7.27</td>
<td>Dolomite</td>
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<tr>
<td>AH02</td>
<td>Madden_AH02</td>
<td>0.14</td>
<td>−7.31</td>
<td>Dolomite</td>
<td></td>
</tr>
<tr>
<td>PP06</td>
<td>Madden_PP06</td>
<td>0.48</td>
<td>−7.43</td>
<td>Dolomite</td>
<td></td>
</tr>
<tr>
<td>AF06</td>
<td>Madden_AF06</td>
<td>0.49</td>
<td>−6.82</td>
<td>Dolomite</td>
<td></td>
</tr>
<tr>
<td>PC08</td>
<td>Madden_PC08</td>
<td>0.60</td>
<td>−5.99</td>
<td>Dolomite</td>
<td></td>
</tr>
<tr>
<td>AK12</td>
<td>Madden_AK12</td>
<td>1.04</td>
<td>−7.48</td>
<td>Dolomite</td>
<td></td>
</tr>
</tbody>
</table>

Granular Mosaic to Blocky Calcite (C1).—C1 calcite cement, retaining “ghost” internal and external fabrics of aragonitic bioclasts, is indicative of neomorphism rather than of bioclast dissolution and reprecipitation. Crosscutting of micritic rims and partial replacement of cone-in-cone cement (Fig. 5A): (1) confirms micritization and cone-in-cone cementation as earlier processes, and (2) is indicative of contemporaneous neomorphic replacement and growth of pore-filling cement. Neomorphism and contemporaneous cement growth (calcitization) into original porosity is consistent with the granular mosaic habit and absence of drusy cement morphologies (Fig. 10; Hendry et al. 1999). A change from dull luminescence to nonluminescence in neomorphic C1 to dull luminescence to bright luminescence in contemporaneous pore-filling C1 blocky cement in coral-hosted microenvironments is consistent with either: (1) trapping of initially oxidizing neomorphic diagenetic fluids that become more reducing within the pores, or (2) that intra-pore porosity received lower amounts and/or less sustained flushing by diagenetic fluids with a resulting decrease in oxygenation state over time. If precipitated from marine parent fluids, cements with stable-isotope values ranging from −6.4 to −11.7‰ δ18O V-PDB, would require temperatures of 40–80 °C, at a burial depth of approximately 500–1700 m. These temperatures and depths are highly unlikely due to the pre-fracture timing of the C1 cement and potentially still oxidizing precipitation conditions. A formation fluid of −7 to −5‰ V-SMOW at 25 to 35 °C is likely (Fig. 9), consistent with early meteoric to shallow-burial diagenesis and similar to the conditions inferred for the cone-in-cone cement that this C1 cement partially replaces. However, samples with low δ18O values may be indicative of: (1) mixing marine pore fluids with meteoric water, or (2) meteoric diagenesis at higher rock:water ratios. With the majority of δ18O values being more negative than −4‰ δ18O V-PDB, these predominantly reflect an influx of meteoric fluids and lower rock:water ratios. High δ18O values (−1.6 to 0.9‰) indicate a lack of soil zone processes affecting these meteoric fluids or a rock-derived source of carbon with marine δ13C values, consistent with the prior cone-in-cone cement.

Aragonitic scleractinian coral skeletons contain ~ 8000 ppm Sr2+ (Milliman 1974; Morrison and Brand 1986). Observations of Sr2+...
concentrations of 3039 to 4692 ppm for C1 neomorphic calcite (Table 2), and assuming a distribution coefficient for Sr of 0.4 or less (Veizer 1983), suggest Sr removal via an open system, but that inflowing fluids contributed little to the elemental signature (Hendry et al. 1999). The high Sr values of C1 may reflect a partially closed system with less effective elemental removal, perhaps within the coral-hosted microenvironments. Scleractinian-coral aragonite typically contains Mn and Fe element concentrations on the order of tens of ppm at most (Morrison and Brand 1986). Low concentrations of Mn (<270 ppm) and Fe (typically <70 ppm) in neomorphic calcite and partition coefficients for both elements of >1 (i.e., ready ion substitution) are again indicative of elemental removal and fluid throughflow, similar to that inferred for the Sr above. These low Mn and Fe concentrations likely are reflected in the dull-luminescent or nonluminescent CL character of the C1 cements.

Intermediate C1 calcite is affiliated with mixed micrite–siliciclastic matrix, which includes Mg/Ca ratios higher than those of C1. In the tropical marine setting of the patch reefs, it is likely that the primary mineralogy of the micrite was either aragonite or high-Mg calcite (Hoskin 1968; Husseini and Matthews 1972; Andrews et al. 1997). The intermediate-Mg calcites may have retained, in part, the precursor elemental composition of the micrite after alteration by undersaturated fluids, or siliciclastic clays might locally contribute magnesium to the elemental signature (Morad et al. 2000).

**Interpretation of Syn-Fracture Features**

Compaction and Fracturing (F1).—The first stage of grain breakage and fracturing (F1) is present mainly in neomorphosed coral-rich bioclastic samples and is interpreted to reflect differential compaction and fracturing of partially lithified matrix, combined with fracturing of more competent cemented components, most notably those units rich in corals (Fig. 10). The timing of fracturing, prior to the onset of later-stage chemical compaction, is also consistent with F1 fractures resulting from mechanical compaction in a shallow burial environment.

Blocky to Equant Cement (C2).—Blocky to equant C2 calcite cement infills, and therefore must post-date, F1 fractures. C2 cements also are present in areas of micritic matrix close to bioclasts where differential compaction has generated porosity, and with some cementation attributable to aggrading neomorphism of the matrix, as suggested by micritic inclusions within the cement (Fig. 5C). A dull-luminescent to nonluminescent character and largely ferroan to weakly ferroan mineralogy indicates that C2 cement probably was precipitated from oxygenated fluids with Fe ions in solution. A positive shift of δ¹⁸O between cone-in-cone calcite (~8.5 to ~10.2‰) and C1 calcite (~6.4 to ~11.7‰) to C2 calcite (~3.6 to ~6.3‰) suggests either: (1) increasing pore-fluid δ¹⁸O, or (2) decreasing temperature. Formation fluid δ¹⁸O values could range from ~0.9 to +1.0‰ V-SMOW at 35–45 °C under conditions of increasing burial (Fig. 9), consistent with a marine source. If precipitated from meteoric fluids, cements with stable-isotope values ranging from ~3.6 to ~6.3‰ δ¹⁸O V-PDB would require temperatures of <20 °C at near-surface burial depths. A burial diagenetic environment is inferred for the C2 cements as evidenced by the prior onset of mechanical fracturing, and indicates that a decreasing formation temperature from C1 to C2 cements is unlikely. Enrichment of the precipitating fluids towards higher δ¹⁸O can be ascribed to: (1) silicate alteration reactions between pore fluids and siliciclastics (Morad 1998), or (2) the introduction of seawater-derived fluids. The depleted δ¹³C values (~3.6 to ~6.3‰ δ¹³C V-PDB) are consistent with carbon derivation from the decay of C3 and C4 plants in soil-rich and sediment-rich horizons (Morad 1998), such as occurs in the associated deltaic siliciclastics. The introduction of Fe ions in solution is most reasonably attributed to fluids sourcing Fe ions from siliciclastics along the fluid flow pathway (Morad et al. 2000).

Compaction and Fracturing (F2 + F3).—F2 and F3 fractures generally are straighter than the earlier F1 fractures (Fig. 5C, D), indicating that fracturing continued after full lithification of matrix and bioclasts. The conjugate natures of some of the F2 and F3 fracture sets (Fig. 5D) conform to either a compaction or a tectonic origin.

Blocky to Equant Cement (C3).—Blocky to equant C3 cementation is confined to, and post-dates, F2 and F3 fractures (Fig. 10). The nonluminescence and mostly nonferroan composition of this cement are consistent with oxygenated diageneric fluids. Stable-isotope values ranging from ~7.3‰ to ~10.7‰ δ¹³C V-PDB would require precipitation at 40–80 °C from marine fluids at a depth of 500–1400 m (based on a regional geothermal gradient of 27.5 °C/km; Hall 2002b), inconsistent with C3 formation prior to the onset of chemical fracturing. Formation-fluid δ¹⁸O values range from ~5.5 to ~2.9‰ V-SMOW at 35–50 °C (Fig. 9) at <1000 m burial depth. At the higher end of this temperature range, formation-fluids for C3 calcite are consistent with precipitation from either marine or meteoric fluids, but lower temperatures and a
nonluminescent CL character suggest mixed water compositions. A shift towards higher δ13C values from C2 to C3 (−1.8 to −0.2‰) cements suggest enrichment of pore fluids by: (1) marine rock-sourced carbon or (2) a lack of soil-zone signatures. An influx of meteoric fluids dissolving marine carbon along its flow path and mixing with residual waters from C2 precipitation is likely with fluids utilizing F2 and F3 fractures as flow pathways.

Fracturing (F4).—The final stage of fracturing is minor and poorly developed, likely due to the impact of prior cementation (cf. Purser 1978; Hird and Tucker 1988; Clari and Martire 1996; Railsback 1993a).

Interpretation of Post-Fracture Features

Stylolites and Dissolution Seams.—Crosscutting all calcite cements and fractures, stylolites and dissolution seams are chemical compaction features that form in moderate to deep burial environments (Machel 2004) at depths of 500–1000 m (where the onset of stylolitization occurs; Nicolaides and Wallace 1997). A transition from dissolution seams in more siliciclastic-rich samples to stylolites in more calcitic, well-cemented samples as occurs in the Samarinda deposits is a trend well documented in the literature (Wanless 1979; Railsback 1993b). The seams mostly circumvent calcified bioclasts, likely because dissolution was promoted along calcite to clay-rich interfaces (cf. Railsback 1993b).

Dolomite Cement.—Dolomites are a localized feature found dominantly in areas of micritic matrix (Fig. 10). These cements commonly are proximal to dissolution seams in the clay-rich matrix or hosted within coral chambers, with Mg perhaps locally contributed from the compaction of clays (Ali 1995; Cornell and Wilson 2004; Machel 2004). The euhedral microcrystalline and partly cloudy dolomite crystals indicate that they are at least in part replacing the micritic matrix (Warren 2000; Machel 2004). Euhedral rhomboic dolomite is indicative of formation temperatures of <50–60 °C (Warren 2000). δ18O values for dolomite (−5.99 to −7.48‰ V-PDB) at 53 to 60 °C convert to formation fluids of −3.4 to −2.3‰ V-SMOW (Fig. 9). On the basis of the combined petrographic and stable-isotope evidence, meteoric fluids at moderate burial depths up to ~1000 m are interpreted to be the dolomitizing fluids; however, slightly higher δ18O values may be the result of: (1) mixed marine–meteoric signature similar to C3 fluids, or (2) silicate alteration reactions between pore fluids and siliciclastics (Morad 1998).

DISCUSSION

Basin Hydrology

Fluids are integral to the diagenesis of sediments, with three primary mechanisms recognized as driving forces for fluid flow through sedimentary successions (Galloway 1984): (1) topography-driven flow, (2) compaction-driven flow, and (3) thermal-convection-driven flow.
Given the conditions of shallow burial (< 1 km; based on the onset of stylolitization) and low thermal heat flow (Hall 2002b) for the Samarinda patch reefs, thermally driven fluid flow is not considered an important factor. Topography-driven flow (established by the elevation of the ground-water table above sea level; Bjørlykke 1993), and compaction-driven flow of trapped marine fluids, appear the more important controls on subsurface flow in the Samarinda area (Fig. 11).

Low oxygen isotope values ($\delta^{18}O$ values of $-6.4$ to $-11.7\%$ V-PDB) suggest that, at low temperatures, meteoric fluids with an upland source were responsible for the formation of early cone-in-cone and neomorphic C1 cements. Involvement of continent-derived fluids in the early diagenesis of the Samarinda patch reefs is consistent with their development adjacent to permeable sands of the Mahakam Delta (Allen and Chambers 1998). The Samarinda patch reefs are underlain by deep-

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**Fig. 8.** — Paragenetic sequence of the patch reefs of the Mahakam Delta. Relative timing of events is on the basis of petrographic observations.

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**Fig. 9.** — Equilibrium oxygen isotope fractionation relationships between calcite and dolomite cements and formation fluids for the Samarinda patch reefs. Calcite curves are based on the equation $10^3 \ln x = 2.78 \times 10^6 T^{-2} - 2.89$ (where $x$ is the fractionation factor of calcite-water and $T$ is temperature in K; Friedman and O’Neil 1977) and the dolomite curve is based on the equation $10^3 \ln x = 3.20 \times 10^6 T^{-2} - 3.30$ (where $x$ is the fractionation factor of dolomite-water and $T$ is temperature in K; Land 1983). For each cement phase, averaged $\delta^{18}O$ V-PDB compositions were used. CC indicates cone-in-cone cement. C2 cements suggest precipitation from pore fluids with increasing $\delta^{18}O$, or precipitation from meteoric fluids at decreased temperatures than both earlier (CC and C1) and later (C3 and D1) cements.
water shales and overlain by alluvial flood-plain sediments (Fig. 2). The location of the reefs adjacent to an “ever-wet” elevated continental landmass is inferred to have contributed to the development of a confined paleo-aquifer allowing channelling of continent-derived fluids laterally into the patch reefs, dissolving marine carbonate along the flow path, and contributing to the high carbon isotope values of the patch-reef cements (Fig. 11). Furthermore, an aquifer system in an “ever-wet” climate would allow replenishment of meteoric fluids capable of pervasive stabilization of the patch reefs through neomorphic replacement and calcitization. Permeable deltaic units have been shown to be important in the introduction of meteoric diagenetic fluids into carbonates during early diagenesis (Hendry et al. 1999; Taberner et al. 2002). Later cementation, fracturing, and compaction are relatively minor features occurring under the influence of continued burial, rock–water interaction, and the confined aquifer system generated by covering of the patch reefs in flood-plain sediments (Fig. 2). Under conditions of increasing burial, the introduction of δ18O-enriched, δ13C-depleted fluids and Fe ions in solution is consistent with dewatering of marine fluids from the underlying deep-water shales possessing an evolved δ18O seawater character sourcing Fe ions from the shales (Fig. 11; cf. Galloway 1984; Hendry et al. 1999).

The diagenesis of a number of SE Asian siliciclastic-influenced coastal carbonates show pervasive early neomorphism of aragonite (Netherwood and Wight 1992; Madden 2008; Wilson 2012), and hint that, similar to the Samarinda patch reefs, their diagenesis is strongly controlled by the role of a continental aquifer driving fluid flow. Few detailed diagenetic studies exist of delta-associated carbonates that developed during the Cenozoic, in humid, temperate, or arid settings. However similarities, and differences, are exhibited between the Samarinda carbonates and delta-associated carbonates from a range of climatic settings from the Cenozoic.

FIG. 10.—Schematic paragenetic scheme from different areas of a patch reef, based on petrographic relationships. Reef margin shows diagenesis of a platy coral-rich section from a mixed carbonate–siliciclastic (35–80% siliciclastics) area of the reef with development of cone-in-cone features, dolomites in the siliciclastic-rich matrix, and dissolution seams. Reef core shows diagenesis of a head or branching coral-rich section containing < 35% siliciclastics in which neomorphic C1 granular to blocky cement is prevalent and stylolites are developed.
Patterns of Diagenesis of Coastal Carbonates in the Humid Tropics

With respect to early diagenesis and initial basin hydrology, the Samarinda carbonates are sedimentologically and diagenetically comparable to the Eocene Calders reef of the South Pyrenean foreland basin. The Calders reef developed on delta-lobes in a warm and humid climatic setting and is the subject of one of the few other detailed diagenetic studies of these types of deposits (Hendry et al. 1999). Reef development occurred contemporaneously with siliciclastic influx, resulting in both admixing and interbedding of mixed carbonate-clastic and siliciclastic lithologies (Hendry et al. 1999; Burton 2003; Lokier et al. 2009). Reef demise is attributed to influx of siliciclastic sediments, now forming low-permeability strata that overlie the Calders reef (Hendry et al. 1999). The Calders reef, like the Samarinda carbonates, was stabilized pervasively early in its diagenetic history via meteoric phreatic to shallow burial diagenesis under the influence of continent-derived fluids in a confined aquifer system (Hendry et al. 1999). Continental waters with an upland source, indicated by \( \delta^{18}O \) SMOW values of \(-7.9\) to \(-9.0\)%o, were channeled through the reef by adjacent siliciclastic units and the low-permeability overlying strata. High \( \delta^{13}C \) values indicate that marine bicarbonate was sourced from aragonite dissolution along the meteoric-water flow path. This flow was, however, not influenced by the soil zone (Hendry et al. 1999). This pervasive early stabilization and cementation, linked to aquifer flow, appear to be essential elements of conceptual models for diagenesis of humid tropical coastal carbonates (Fig. 11; cf. Moore 2001). Nonetheless, the late diagenetic history of the Calders reef differs markedly from the Samarinda patch reefs associated with a postdepositional shift in climate towards more seasonal and arid conditions (Cavagnotto and Anadón 1996). Limited aquifer recharge resulting from an arid climatic shift facilitated late-stage cementation by marine fluids driven off from compacting basinal marls and evaporites at Calders (Hendry et al. 1999; Taberner et al. 2002).

Comparisons with Humid, Temperate Coastal Carbonates

Pliocene limestones (the Te Aute Formation) from the eastern North Island of New Zealand are nontropical, cool-water, skeletal carbonate deposits (Nelson et al. 2003; Caron et al. 2005). Perhaps paradoxically, these coastal cool-water limestones provide further analogs to the Samarinda carbonates; the Te Aute Formation developed as near-shore shelf carbonates in a predominantly siliciclastic setting adjacent to a large elevated continental landmass. The limestones of the Te Aute Formation developed in two settings: (1) under strong, high-energy tidal flow, and (2) with weaker tidal currents, resulting in fine-grained siliciclastic-carbonate admixtures (Beu 1995). The primary low-Mg calcite mineralogy of the cool-water skeletal carbonates contrasts with the more aragonite-rich tropical Samarinda carbonates. However, aragonite locally constitutes 10–60% of the limestones (Hood and Nelson 1996; Nelson et al. 2003; Caron and Nelson 2009). Differential tilting and uplift of the limestones in basin-margin areas allowed temporally and spatially variable contact between the limestone and recharging meteoric waters (Kamp and Nelson 1988; Dodd and Nelson 1998; Nelson et al. 2003). Despite this variable influence of meteoric fluids, and the dominance of metastable low-Mg calcite mineralogy at Te Aute (Caron and Nelson 2009), the aragonite-rich sections of the Te Aute Formation were strongly affected by selective meteoric dissolution, and, to a lesser degree, neomorphism of aragonite provided a shallow source of carbonate for calcite cements with depleted (meteoric) stable-isotope composition. In comparison, the Samarinda carbonates were stabilized and calcitized by meteoric groundwater fluid flux providing carbonate in solution with a rock-derived, rather than a soil-zone-derived, carbon isotope signature.

Comparisons with Subtropical Arid Coastal Carbonates

Although subtropical arid coastal carbonates are well known from areas such as the Mediterranean and the Red Sea, the diagenesis of these carbonates is not as extensively studied as their sedimentology or stratigraphy (Esteban 1996). The diagenetic characteristics of these arid systems show major differences from humid tropical coastal deposits. Upper Miocene deposits of the Mediterranean region and middle Miocene strata of the Red Sea include subtropical arid carbonates (Hayward 1982; Burchette 1988; Santisteban and Taberner 1988; Sun and Esteban 1994; Esteban 1996; Burton 2003). Many of these carbonates developed on marginal-marine high-energy fan-delta systems, resulting in intercalated wedges of pure carbonates and siliciclastic sediments due to periodic flash-flood-related influx of terrigenous siliciclastics (Santisteban and Taberner 1988; Wilson and Lokier 2002; Burton 2003). Poor sorting of the intercalated siliciclastics results in limited development of high-
permeability aquifers. In contrast with humid equatorial carbonates, the diagenesis of subtropical arid carbonate systems commonly is influenced by flushing of hypersaline brines, leading to extensive dolomitization (Sun 1992; Sun and Esteban 1994; Esteban 1996). Carbonates developed in the semiarid climate of the lower and middle Miocene of the Mediterranean show evidence for diagenetic alteration by meteoric fluids. However, unlike the pervasive early stabilization evident in the equatorial humid coastal systems, extensive meteoric leaching and complete dissolution of skeletal aragonite occurs in pure carbonates associated with fan-delta sediments (Burton 2003). Drusy meteoric cements may partially fill primary and secondary porosity, and preserved porosity may be as high as 20–55% (Burton 2003). The Samarinda carbonates are extensively stabilized and cemented by aquifer flow. The development of subtropical carbonates on, and encasement by, poor-aquifer-quality units likely results in nonpervasive and poorly developed meteoric groundwater flow paths, especially when coupled with an arid climatic setting. Even with preferred flow units, due to the decreased volumes of water flux in subtropical arid coastal carbonates, meteoric aquifer-related stabilization of aragonite is limited and there is good potential for porosity preservation due to limited cementation (cf. James and Choquette 1990; Sun and Esteban 1994).

Conceptual Diagenetic Model of Delta-Associated Carbonates in Humid Tropical Settings

The interpretation of the diagenetic history of patch reefs of this study, as with that of other warm, humid tropical coastal examples (Hendry et al. 1999), suggests a model of continental aquifer flow driving early pervasive stabilization (Fig. 11). For the Samarinda carbonates, the lack of subaerial exposure due to early burial by siliciclastics, the presence of permeable siliciclastic units in association with the reefs adjacent to a large landmass, and the humid equatorial climate are the factors contributing to stabilization and pervasive cementation of the reefs. Other carbonates developed in similar humid coastal settings with high siliciclastic influx and, most importantly, continental-aquifer influence also include evidence of pervasive stabilization and calcitization (Netherwood and Wright 1992; Ali 1995; Hook and Wilson 2003; Wilson 2012). Factors that result in less pervasive stabilization in humid coastal carbonates are: (1) high original primary porosity linked to high depositional energy, (2) more common calcitic components, and (3) localized or limited aquifer flow (Hood and Nelson 1996; Nelson et al. 2003; Caron and Nelson 2009).

CONCLUSIONS

Diagenesis.—Study of Neogene siliciclastic-influenced patch reefs in Borneo reveals that both pervasive early stabilization of reef components from aragonite to low-Mg calcite and calcite cementation are key diagenetic features of these coastal carbonates from the humid tropics. The early stabilization of reef components and extensive cementation are attributed to the reefs being flushed by primarily undersaturated meteoric water. Depleted oxygen stable-isotope data indicate a continental meteoric, upland source for the precipitating fluids with a lesser influence of meteoric to shallow-burial environments. With the exception of an initial phase of grain micritization under marine conditions. Late-stage diagenetic processes of fracturing and cementation (also related to aquifer-driven flow) are inferred to have occurred at burial depths of < 1000 m and at temperatures of up to 53 °C prior to chemical compaction. Overall, the diagenesis of the Samarinda patch reefs is inferred to be controlled dominantly by fluid flow from a continental aquifer, one not directly influenced by “soil-zone” processes (as indicated by generally high δ13C), that caused early stabilization and cementation.

Controls on Diagenesis.—The diagenesis of these patch reefs from equatorial “ever-wet” climatic settings are strongly controlled by basin-margin paleohydrology. Notably, continental aquifer flow, which drove shallow subsurface flow and early pervasive stabilization of reef components, was driven by the humid “ever-wet” climatic conditions associated with extensive Cenozoic tectonic uplift. Early stabilization and cementation are developed in other carbonates in similar humid mixed carbonate-siliciclastic successions, with large adjacent landmasses, in which extensive meteoric phreatic to shallow-burial fluid flow is the dominant diagenetic process. These findings are of particular relevance to SE Asia, with widespread mixed carbonate-siliciclastic systems developed in coastal settings with a humid equatorial climate, in which models generated in other regions cannot easily be applied.

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