The mineralogic and isotopic fingerprint of equatorial carbonates: Kepulauan Seribu, Indonesia

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Abstract
Kepulauan Seribu is an isolated patch reef complex situated in the Java Sea (Indonesia) and is a typical example for a humid, equatorial carbonate system. We investigate the mineralogical and isotopic fingerprint of Panggang, one of the reef platforms of Kepulauan Seribu, to evaluate differences to other carbonate systems, using isotope in combination with XRD and SEM analysis. A characteristic property of shallow water (< 20 m) sediments from Kepulauan Seribu is their increased LMC content (~ 10%) derived from some genera of rotaliid foraminifers and bivalves. The relative abundance of these faunal elements in shallow waters might be related to at least temporary turbid conditions caused by sediment-laden river runoff. This influence is also evidenced by the presence of low amounts of siliciclastic minerals below the regional wave base. Kepulauan Seribu carbonates are characterized by very low δ13C and δ18O values. This is related to the isotopically depleted riverine input. The δ13CDIC in riverine water is reduced by the contribution of 12C from riverside mangroves. Deep atmospheric convection and intensive rains contribute 18O-depleted freshwater in the river catchments, finally reducing salinity in the Java Sea. The depleted δ13C signature in carbonates is further enhanced by the lack of green algae and inorganic carbonates and abundance of coral debris. Low δ18O values in carbonates are favored by the high water temperatures in the equatorial setting. Since equatorial carbonates in SE Asia, including the Java Sea, are typically influenced by high turbidity and/or river runoff, the observed distinctively low isotope values likely are characteristic for equatorial carbonate systems in the region.

Keywords Patch reefs · Oxygen isotopes · Carbon isotopes · SE Asia · Humid equatorial climate · Maritime continent

Introduction
During the last decade the awareness has grown that carbonate systems in the humid equatorial zone have common characteristics that distinguishes them from carbonates in more arid regions (Tomascik et al. 1997; Wilson 2008). Features of these equatorial carbonates include the common occurrence of photozoan assemblages adapted to oligophotic (low light) conditions, common elements of the Heterozoan Association and mixed carbonate/clastic deposits. Humid equatorial carbonates are also characterized by a lack of coated grains, aggregates, chemically precipitated mud and the absence of associated evaporites (Wilson 2002). They typically occur in a zone between 15/20° N and 15/20° S in areas with warm (> 22 °C) surface waters and heavy precipitation (> 1500 mm per annum; Wilson, 2012). Due to the increased precipitation, carbonate deposition takes place in shelfal waters of normal to reduced salinity and high clastic influx. River runoff and upwelling cause increased nutrient concentrations in surface waters, which in turn lead to a shallow position of the photic zone.

A key region for the study of these humid equatorial carbonates is SE Asia, where the shallow shelves are strongly affected by river runoff and/or upwelling (Tomascik et al. 1997; Wilson 2002). The diversity of environments and habitats renders SE Asia an ideal natural laboratory to investigate the variability of equatorial carbonate processes.
and deposits (Tomascik et al. 1997; Wilson 2002, 2008). Sediment discharge from each of the four largest SE Asian islands varies from ca 300 to 1650 million tons/year (Milliman et al. 1999), leading to high turbidity and nutrient concentration in surface waters. Depth of abundant coral growth in SE Asian reefs is therefore limited to approximately 20–30 m and is considered among the shallowest in the world (Schlager 1992; Wilson 2008). This shallow photic zone corresponds with low light penetration in an oligo-mesotrophic region characterized by nutrient input from river runoff, upwelling and plankton blooms (Wilson 2008). Due to these environmental conditions, SE Asian carbonates can be considered as typical equatorial carbonate systems.

The low salinity in the semi-enclosed seas of SE Asia also leads to a relatively low aragonite supersaturation compared to more arid regions, such as the tropical, western Atlantic (TWA). The development of Cenozoic carbonate systems in SE Asia is thought to be strongly influenced by regional changes in nutrient input and precipitation (Wilson 2008). Late Oligocene carbonates from SE Asia were often characterised by calcitic dominated assemblages. This was likely promoted by the decreased aragonite supersaturation of the low salinity waters in the semi-enclosed seas of SE Asia (Wilson 2012). An increase of river runoff and associated decrease in aragonite supersaturation is also held responsible for a shelf wide switch from inorganic aragonite to bioclastic calcite dominated sedimentation on the tropical NW Shelf of Australia during the early Holocene (Hallenberger et al. 2019; Ishiwa et al. 2019); a process which seems to have occurred repeatedly during the Pleistocene (Gallagher et al. 2018). Studies of modern carbonate systems from northern Australia and SE Asia suggest that faunal assemblages from the oligophotic zone are relatively calcite-rich (Wilson and Vecsei 2005; Wilson 2012). Generally, this change from coral dominated euphotic to foramol dominated oligophotic zones takes place in a water depth of around 20 m, but locally can be as shallow as 6 m in turbid coastal water (Wilson and Vecsei 2005).

Equatorial carbonates share some similarities with cool water systems including the abundance of calcitic components and heterozoan elements. In some cases, this can complicate the distinction between non-tropical and equatorial carbonates (Wilson and Vecsei 2005). Wilson (2012), therefore, recommended the use of additional geochemical and isotopic proxies to identify warm temperatures and low salinities associated with regional equatorial carbonates in the rock record. However, stable isotope studies from modern SE Asia that could be used as a template for comparison with fossil counterparts are still missing. It is also unclear if the predominance of calcitic bioclasts, that is apparent in the oligophotic zone of many carbonate systems in SE Asia, is visible in the mineralogy of the coral dominated shallow water area. In this study, we investigate how far the mineralogy and stable isotope signature of shallow water (<20 m) equatorial carbonate from Kepulauan Seribu in the Java Sea is distinct from other carbonate systems. Stable isotope (δ13C and δ18O) analysis in combination with X-ray diffraction (XRD) and Scanning Electron Microscopy (SEM) is used to provide a fingerprint for equatorial carbonates from SE Asia. This data might be used as a template for comparison with carbonates in the rock record.

**Study area**

Situated in the Java Sea, Kepulauan Seribu (The Thousand Islands) consists of numerous coral reef islands ranging in size from a few meters across to length greater than one kilometer. From about 10 km off the northwest Java coastline, the island chain extends for about 50 km into the Java Sea. Kepulauan Seribu is situated on the NNE-SSW structural high of the Seribu platform, which is separated from the Sunda Basin towards the west and Arjuna basin to the east by the active N-S oriented Seribu Fault system (Scrutton 1976a). The north–south elongation of the platform is largely controlled by the regional fault pattern that was established during the early Cenozoic and reactivated at various times since (Park et al. 2010). The patch reef system of Kepulauan Seribu is located on the shallow Sunda Shelf where it rises from a water depth of 30–50 m to the sea surface. A deep and several kilometers wide, east–west oriented channel separates Pari Island to the south from most other islands to its north (Fig. 1). It appears that this channel forms an important hydrologic barrier, channeling the sediment-laden coastal waters off Java and Kalimantan away from the reef (Tomascik et al. 1997; Jordan 1998). The reefal buildups on the platform vary in size and shape but most of them are elongated in a roughly east–west orientation. This pronounced orientation is inherited from the Pleistocene drainage pattern which was abandoned and truncated by river capture prior to the Holocene transgression (Park et al. 2010). The islands of Kepulauan Seribu are considered to be located in a relatively sheltered position protected from severe storms and ocean swells due to the fact that it is surrounded by major landmasses, such as Sumatra, Java and Kalimantan (Fig. 1).

The controlling influence on modern reef growth and morphology in Kepulauan Seribu is believed to be the seasonal change in wind and current directions (Scrutton 1976a). The Java Sea is characterized by a monsoonal climate with a seasonal reversal between the West Monsoon (December–February), and the East Monsoon (April–October). The West Monsoon winds are known to be stronger than the East Monsoon winds, but less persistent (3 vs. 7 months). The East Monsoon winds therefore are the dominant factor for the net sediment transport (Poirbandono...
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2016). The climate in Kepulauan Seribu is tropical with relative humidity of 80–90% (Tomascik et al. 1997). The maximum sea surface temperature (SST) is observed in April (29.8 °C) and minimum in January (28.6 °C) (Cahyarini et al. 2016). Surface salinity in the Java Sea (30–34‰) is well below normal open ocean levels, reaching as low as 22‰ in the vicinity of some of the larger river mouths. The low salinity of the Java Sea is due to heavy precipitation and intensive river runoff from Java and Kalimantan (Sofian and Kozai 2007). The water column of the shallow western Java Sea is well mixed and shows little temperature or salinity stratification (Hamzah et al. 2020).

Water along Kepulauan Seribu is highly eutrophic (Baum et al. 2015). Seasonal variability of Chlorophyll-α distribution in the Java Sea is influenced by monsoon, through water exchange with adjacent seas and nutrient supply from river discharge (Koropitan and Ikeda 2016). In the eastern Java Sea, episodic upwelling leads to phytoplankton blooms during the East Monsoon (Koropitan and Ikeda 2016). The role of riverine nutrient input during the West Monsoon (rainy season) is limited to coastal regions (Koropitan and Ikeda 2016). The reefs in Kepulauan Seribu are directly influenced by intense anthropogenic pressure from local as well as regional sources (Baum et al. 2015). Polluted water masses from the Greater Jakarta Area are considered to affect reefs in nearshore areas close to river mouths around Jakarta Bay, while localized effects of anthropogenic stressors are more important for reef in the central and northern reefs of Kepulauan Seribu (Damar et al. 2012; Farhan and Lim 2012; Baum et al. 2015).

Each island and platform of Kepulauan Seribu is built up from the skeletal debris of coral reefs which rim the platform margins (Jordan 1998). The steep reef front extends seaward from the reef crest to a depth of about 15 m. Well-developed reef facies occur down to a depth of about 16–18 m (Jordan 1998), whereas reef development is restricted to an even shallower depth of 3–10 m around 40 km towards the south in Jakarta Bay (Baum et al. 2015). Reef framework builders at Kepulauan Seribu consist of a diverse fauna of Indo-Pacific corals and encrusting red algae. Below the base of continuous reef development, occurs a transition zone of small reef knolls or coralgal knobs that are surrounded by unconsolidated sediments. In water depths of about 24 m, the seafloor becomes entirely sediment and begins to slope less steeply. Below the reef and near reef zone, in a depth from 27 to 88 m occurs a foraminiferal molluscan packstone facies that is rich in terrigenous mud. The dominant bioclasts are benthic foraminifers, mollusc and coral fragments.

**Fig. 1** a Map of Indonesia, surrounding countries, and the study area (blue box). b Bathymetry in the vicinity of Kepulauan Seribu (basemap from National Bathymetry, Indonesia Geospatial Information Agency. Available on http://tides.big.go.id/DEMNAS/). Location of Panggang is indicated by the blue box, and Bidadari is indicated by the red box. c Satellite image showing Panggang reef platform. Sampling spots are indicated as red dots (satellite image from Google Earth).
Statistical analyses based on texture and composition reveal four sedimentary facies in the shallow waters (<20 m) of the Kepulauan Seribu: coral grainstone, coral packstone/grainstone, coral-mollusc packstone and mollusc wackestone (Utami et al. 2018). The occurrence of mollusc wackestone in the lagoon is controlled by water depth, while the sand apron and reef front do not show significant facies separation with water depth. Environmental facies maps were derived from remote sensing data (Utami et al. 2018): The sand apron can be subdivided into three facies with either no, sparse, or dense seagrass cover. One facies is further characterized by active coral growth. The deeper water zone can be separated into a shallow and a deep subtidal part of lagoons and platform margins. Distribution is therefore mainly controlled by water depth, density of seagrass cover and coral abundance (Utami et al. 2018). The satellite-derived environmental facies correlated only in the lagoon directly with sedimentary facies. No direct correlation of environmental facies correlated only in the lagoon directly with sedimentary facies was weighed. Mean grain size and sorting were calculated using the software Gradistat (Blott and Pye 2001). Point counting analysis was done on the split of the >125 µm fraction by counting 300 grains under a binocular microscope. The results of these analyses were reported previously in Utami et al. (2018). Here, we present data on the mineralogy, texture and isotopic composition of the same sample set.

Mineralogy was analyzed using a Bruker D8 X-ray diffractometer equipped with a Cu-anode, a secondary graphite monochromator and a scintillation counter. Forty-five bulk surface sediment samples covering all sedimentary facies (Utami et al. 2018) were oven-dried, ground and mounted on sample holders. Additionally, the mineralogy of the <125 µm fraction from five of these samples and five individual bivalve shells was measured separately. All selected bivalve shells were clean and devoid of incrustations or epibenthos. All measurements were conducted over an angular field of 48° (16°–64°) with a step size of 0.02° per five seconds. Identification and quantification of different mineral phases was achieved by standard Rietveld refinement using the software Profex (ver. 3.14.0). Standard analytical error is ±2%. A total of 20 weight percent of corundum was added to each sample as an internal standard. This standard was systematically overestimated during Rietveld refinement (avg. 4%), which was attributed to the occurrence of amorphous silica in the form of sponge needles. The MgCO₃ content of calcite (in mol %) was determined using the shift in the position of the d(104) peak (Lumsden 1979). The standard error is ±0.15% on the mean composition.

Measurements of δ¹³C and δ¹⁸O values on 15 bulk and five fine fraction samples were made with a Finnigan MAT 253 mass spectrometer connected to a Kiel IV carbonate preparation device at the Leibniz Laboratory, Kiel University. Additionally, nine individual benthic foraminifers (Calcarina, Amphistegina and Sorites) and five individual bivalve shells were measured separately. Bivalve shells were ground to powder before measurement. All Calcarina and Amphistegina, as well as one individual Sorites were split into two aliquots and measured as subsamples. Standard external error is better than ±0.05‰ for δ¹³C and ±0.08‰ for δ¹⁸O (standards NBS 19 and 20). δ¹³C and δ¹⁸O values are reported in ‰ versus the Vienna PeeDee belemnite (V-PDB). A core of 68 cm length was drilled in September 2005 from a massive Porites coral in the Kepulauan Seribu complex, inshore of Bidadari Island at site 6°01′55″S, 106°44′47″E (Fig. 1) from 3 m water depth (Cahyarini et al. 2016). δ¹³C and δ¹⁸O values were measured at the Stable Isotope Laboratory, Vrije University, Amsterdam. For the analysis, the samples were reacted with 100% H₃PO₄ at 75 °C in a Kiel
Device connected to a Finnigan MAT 252 mass spectrometer. Standard external error is ± 0.05‰ for δ¹³C and ± 0.08‰ for δ¹⁸O (standards NBS 19 and GICS).

A total of 16 selected samples of the < 125 μm fraction were further analysed using scanning electron microscopy in the SEM laboratory of FMIPA (Institut Teknologi Bandung) and the SEM/EDS laboratory of UPP Chevron-Prodi Geologi (Institut Teknologi Bandung). Satellite images were acquired from Google Earth (http://www.google.com/earth). The inverse distance weighting tool of QGIS (ver. 3.14.1-Pi) was used to interpolate data between the sampling points for the generation of maps. All statistical analyses were performed using PAST (Hammer et al. 2001). The Levenés test was used to check if homogeneity of variances was met. The non-parametric Mann–Whitney U test was used for pairwise comparisons of median values. The non-parametric Kruskal–Wallis test for independent samples was used to examine the medians of δ¹³C and δ¹⁸O values in different sedimentary facies. The significance level for all analysis was set at 95%.

Table 1 Average values of grain types and bulk mineralogy for each sedimentary facies (%). Grain types are reported as % of > 125 μm fraction, fines (< 125 μm) as % of total sediment

| Facies       | Coral | Red Algae | Hal | Mollusc | Foram | Echi | N-S | other | Fines | ARA  | HMC | LMC | Qz | Sm | Mg (%) |
|--------------|-------|-----------|-----|---------|-------|------|-----|-------|-------|------|-----|-----|----|----|--------|
| C G          | 58.8  | 5.5       | 1.7 | 19.5    | 8.0   | 4.2  | 1.1 | 1.1   | 6.2   | 80   | 9   | 10  | 1  | 0  | 14.2   |
| C P/G        | 53.7  | 3.4       | 3.5 | 28.2    | 4.9   | 3.9  | 0.3 | 2.2   | 7.0   | 82   | 8   | 9   | 1  | 0  | 14.2   |
| C-M P        | 48.1  | 3.8       | 2.5 | 28.1    | 5.6   | 6.1  | 0.7 | 5.0   | 23.4  | 81   | 9   | 9   | 1  | Trace | 13.8  |
| M W          | 24.0  | 1.3       | 0.8 | 40.7    | 3.6   | 5.7  | 2.4 | 21.5  | 51.5  | 77   | 8   | 11  | 2  | 2  | 13.0   |

C G Coral Grainstone, C P/G Coral Packstone/Grainstone, C-M P Coral-Mollusc Packstone, M W Mollusc Wackestone, Hal Halimeda, Foram foraminifers, Echi echinoderm, N-S non-skeletal, ARA aragonite, HMC high magnesium calcite, LMC low magnesium calcite, Qz quartz, Sm smectite, Mg (%) mole % MgCO₃ in Mg-Calcite

Results

Mineralogy

Sediments in Kepulauan Seribu are mostly composed of biogenic grains, most notably corals and molluscs, with minor components including benthic foraminifers, echinoderms, red algae and the green alga Halimeda. Non-skeletal grains are very rare and restricted to faecal pellets and aggregates cemented by red algae or serpulids (Table 1) (Scrutton 1976b; Utami et al. 2018). SEM investigation confirms that the fine fraction of sediments in Kepulauan Seribu is largely a product of skeletal breakdown (Fig. 2). X-ray diffraction results confirm that surface sediment on the Panggang reef platform in the central part of Kepulauan Seribu is composed mainly of carbonate minerals with small percentages of siliciclastic minerals. Aragonite is the dominant mineral and its contribution varies between 72 and 88%, with a mean of 82% (Table 1, Fig. 3). High magnesium calcite (HMC) and low magnesium calcite (LMC) contents are fairly equal throughout the island, for HMC, the value ranges from 5 to 10%.

Fig. 2 SEM images of the fine fraction: a Sample MPG 11P from the southern reef flat. Sediment constituents are mostly bioclasts formed by skeletal breakdown including calcitic (LMC) prisms from bivalve shells (M). b Sample MPG 41L from the northwestern reef flat. Arrow points at open microborings.
16% with a mean of 8%, for LMC the value varies between 5 and 17% with a mean of 9% (Fig. 4). Quartz composes only a minor fraction of the total bulk sample, with a maximum of 3% and a mean of 1%. From 45 bulk samples tested, three samples show smectite contents between 2 and 9%.

Throughout all sedimentary facies (Utami et al. 2018), the mineralogical composition varies only slightly (Table 1, Fig. 3). The HMC content is very similar in all facies, with an average value of 8.5%. The MgCO₃ content in HMC varies between 13 and 14.2% in mollusc wackestone and coral grainstone facies, respectively. The content of LMC is equal or higher compared to HMC in all facies. The highest aragonite (82%) and lowest LMC contents (9%) are found in the coral packstone/grainstone facies. The mollusc wackestone facies shows the lowest aragonite (77%) and highest LMC (11%) content. The highest siliclastic content is present in the mollusc wackestone facies (quartz: 2%, smectite: 2%).

In addition, the mineralogy of the fine fraction (<125 µm) was measured for five samples (Table 2 and Fig. 5). Compared to their respective bulk samples, the fine fraction shows ~1% higher quartz and significantly higher (9%) smectite contents. If only the carbonate composition is considered, on average, the fine fraction shows lower aragonite (bulk: 79%, <125 µm: 73%), higher HMC (bulk: 10%, <125 µm: 14%) and nearly equal LMC (bulk: 10%, <125 µm: 11%) contents. XRD analysis confirmed that all of the five individual bivalve shells are bimineralic, containing predominantly aragonite (90%) and LMC (10%) as a subordinate phase.

**Stable isotopes of carbonates**

Carbon isotope values (δ¹³C) range from −1.3 to 0.9 ‰ and oxygen isotopes (δ¹⁸O) from -3.4 to -4.9 ‰ for bulk sediment (Fig. 6). The fine fraction (<125 µm) samples did not show a systematic offset in their isotope values with respect to their bulk samples (Fig. 6). Coral abundance is negatively correlated with bulk sediment δ¹³C (r = −0.52; p = 0.04) and δ¹⁸O (r = −0.79; p < 0.01; Table 3), while mollusc abundance shows a positive correlation with δ¹³C (r = 0.58; p = 0.02) and δ¹⁸O (r = 0.79; p < 0.01). A negative correlation is also apparent for the relationship between the abundance of aragonite and δ¹⁸O (r = −0.7; p ≤ 0.01). Aragonite and δ¹³C also show a weak negative correlation (r = −0.43, p = 0.11). Coral isotope values from Bidadari (Fig. 6) are centered around −3 ‰ δ¹³C and −6 ‰ δ¹⁸O and lower compared to the bulk samples. Isotope values from bivalves and benthic
foraminifers are generally higher compared to the bulk samples but partly overlap with samples from the mollusk wackestone facies (Fig. 6). Carbon and oxygen isotope values of bulk sediment from Kepulauan Seribu show more positive value in the inner lagoon and more negative value in the reef front (Fig. 7).

Discussion

Spatial and compositional pattern in mineralogy

The dominant mineral phase in Panggang is aragonite (Table 1). Its concentration varies little between facies
The aragonitic red algae genera *Galaxaura* and *Liagora* also thrive on the Kepulauan Seribu reef slope (Atmadja 1977), likely also contributing to the high aragonite concentrations. HMC and LMC occur in fairly similar amounts in all facies (Table 1, Fig. 3), with slightly increased values of LMC in the mollusc wackestone facies of the inner lagoon and some isolated points on the western reef flat (Fig. 4). Corals as the main aragonite producers are less abundant in this area, while less light dependent carbonate producers, such as molluscs and echinoderms, are more common (Table 1). The highest LMC content is found in the mollusc wackestone facies (Table 1), which is known to dominate in the inner lagoon where coral grainstone and coral packstone/grainstone facies are absent (Utami et al. 2018). High LMC content in the inner lagoon is likely due to the contribution from molluscs, the most important component in the mollusc wackestone facies (Table 1). Bivalves with calcitic or mixed aragonite-calcite shells in Kepulauan Seribu belong to the families Ostreidae (LMC), Pectinidae (LMC + ARA) and Pinnidae (LMC + ARA) (van der Meij et al. 2009; Ueda 2020). Five individual bivalve shells were measured using XRD. All of them have a mixed aragonite–calcite shell containing on average 90% aragonite and 10% LMC. Locally, LMC is also abundant in the reef flat (Fig. 4), likely due to contributions from bivalves and larger benthic foraminifers (e.g. *Amphistegina* and *Elphidium*, Table 4) (Blackmon and Todd 1959) which are abundant in the coral grainstone

**Table 2** Mineralogy (%) of five fine fraction (< 125 µm) samples, along with their respective bulk sample mineralogy

| Sample | Sedimentary facies | C-M P | M W | M W | M W | C-M P |
|--------|--------------------|-------|-----|-----|-----|-------|
| < 125 µm | Carbonate* bulk mineralogy | ARA | 24% | 40% | 71% | 75% | 18% |
| | HMC | 79 | 83 | 77 | 75 | 81 | 79 |
| | LMC | 11 | 8 | 11 | 12 | 9 | 10 |
| | Mg (%) | 9 | 8 | 11 | 13 | 9 | 10 |
| | Carbonate* < 125 µm | ARA | 13.0 | 14.0 | 11.7 | 12.7 | 14.0 | 13.1 |
| | HMC | 74 | 76 | 73 | 71 | 74 | 73 |
| | LMC | 14 | 13 | 14 | 15 | 13 | 14 |
| | Mg (%) | 12 | 9 | 13 | 12 | 10 | 11 |
| | Bulk mineralogy | ARA | 78 | 83 | 73 | 72 | 79 | 77 |
| | HMC | 11 | 8 | 10 | 11 | 9 | 10 |
| | LMC | 9 | 8 | 11 | 13 | 9 | 10 |
| | Qz | 1 | 1 | 2 | 3 | 2 | 2 |
| | Sm | 0 | 0 | 4 | 0 | 0 | 1 |
| < 125 µm | Carbonate* < 125 µm | ARA | 66 | 62 | 64 | 64 | 62 | 64 |
| | HMC | 13 | 11 | 13 | 13 | 11 | 12 |
| | LMC | 11 | 8 | 12 | 11 | 8 | 10 |
| | Qz | 4 | 4 | 3 | 2 | 4 | 3 |
| | Sm | 7 | 14 | 9 | 8 | 11 | 10 |

Carbonate* is mineralogy normalized to only carbonate mineral phases

*Sed facies* sedimentary facies, C-M P coral-mollusc packstone, M W mollusc wackestone, ARA aragonite, HMC high magnesium calcite, LMC low magnesium calcite, Qz quartz, Sm smectite, Mg (%) mole % MgCO₃ in Mg calcite

**Fig. 5** Average mineralogy (%) of five fine fraction (< 125 µm) samples, along with their respective bulk sample mineralogy as indicated in Table 2
facies (Table 1). SEM analysis shows that calcareous nanoplankton, as a potential LMC source, is near absent in the sediment.

No clear correlation exists between HMC and individual biogenic grain types (Table 3). This is caused by the varied contribution of several different biota to the overall HMC content to the sediment. Red algae, such as *Amphiroa*, that are typically associated with sea grass in Kepulauan Seribu (Atmadja 1977), contribute HMC to the coral grainstone facies on the reef flat (Table 1). Echinoderms are more important in the deeper water coral-mollusc packstone facies and mollusc wackestone facies (Table 1) compared to other facies. Increased HMC contents in the inner lagoon (Fig. 4) can therefore be interpreted as a higher contribution from echinoderms. Mg content is generally higher in coralline algae compared to other organisms with HMC skeleton (Chave 1954). This is consistent with the observation that the lowest Mg contents in HMC are found in the mollusc wackestone facies where coralline algae are rare (Table 1, Fig. 3).

The mineralogy of five fine fraction (< 125 \(\mu m\)) subsamples from the coral-mollusc packstone and mollusc wackestone facies was analyzed using XRD. The aragonite content in these fine fraction subsamples is 6% lower compared to their respective bulk samples (Table 2). In contrast, their HMC content is increased by 4% (Table 2). It has been proposed that fish can produce significant amounts of HMC mud (Salter et al. 2012). Most of the important (> 5% abundance) fish families in the area of Kepulauan Seribu (Madduppa 2013) produce Mg-rich amorphous calcium carbonate. This carbonate phase commonly dissolves within hours and therefore is unlikely to be preserved as sediment (Salter et al. 2017). Only one family (Apogoniidae) produces mud-sized HMC ellipsoids with Mg content of around 30–45% (Salter et al. 2018). However, the typical crystal morphologies have not been observed during SEM investigations in our samples. It is, therefore, unlikely that fish-derived Mg calcite contributes significantly to the increased HMC values in the fine fraction. In contrast, SEM investigations confirm that the fine fraction is largely a product of skeletal breakdown (Fig. 2). Carbonate components that occur only in the mud to fine sand size (< 125 \(\mu m\)) fraction, such as tunicate spiculae and needles from Padina thalli, are exclusively aragonitic. Laboratory studies suggest
Table 3 Correlation matrix: Significant correlations ($p < 0.05$) are marked in bold

|           | Coral | R Algae | Hal | Mollusc | Foram | Echi | Fines | ARA  | HMC  | LMC  | Qtz  | Sm  | Water depth | $\delta^{13}C$ | $\delta^{18}O$ |
|-----------|-------|---------|-----|---------|-------|------|-------|------|------|------|------|-----|-------------|----------------|----------------|
| Coral     | 0.051 | 0.34    | <0.01 | 0.15    | 0.32  | <0.01| 0.06  | 0.86 | 0.03 | 0.04 | 0.10 | 0.66 | 0.049 | <0.01 |
| R Algae   | 0.29  | 0.98    | <0.01 | <0.01  | 0.19  | 0.06 | 0.45  | 0.78 | 0.88 | 0.30 | 0.36 | 0.42 | 0.10 | 0.22 |
| Hal       | 0.14  | −0.003  | 0.15 | 0.19    | 0.10  | 0.052| 0.60  | 0.08 | 0.47 | 0.34 | 0.62 | 0.25 | 0.84 | 0.46 |
| Mollusc   | −0.93 | −0.54   | −0.22 | <0.01  | 0.55  | <0.01| 0.10  | 0.66 | 0.09 | 0.08 | 0.18 | 0.92 | 0.02 | <0.01 |
| Foram     | 0.22  | 0.58    | −0.20 | −0.48  | <0.01 | 0.02 | 0.29  | 0.98 | 0.22 | 0.31 | 0.27 | 0.61 | 0.02 | <0.01 |
| Echi      | −0.15 | 0.20    | −0.25 | −0.09  | 0.49  | 0.13 | 0.89  | 0.53 | 0.55 | 0.40 | 0.72 | 0.11 | 0.02 | 0.16 |
| Fines     | −0.74 | −0.28   | −0.29 | 0.74   | −0.33 | 0.23 | <0.01 | 0.55 | 0.01 | <0.01| <0.01| 0.23 | 0.26 | <0.01 |
| ARA       | 0.28  | −0.12   | 0.08 | −0.25  | 0.16  | 0.02 | −0.47 | <0.01| <0.01| <0.01| <0.01| 0.53 | 0.11 | <0.01 |
| HMC       | −0.03 | 0.04    | −0.26 | 0.07   | 0.00 | −0.10 | 0.09  | −0.66 | 0.09 | 0.22 | 0.95 | 0.19 | 0.03 | 0.04 |
| LMC       | −0.32 | 0.02    | 0.11 | 0.26   | −0.18 | 0.09 | 0.41  | −0.78 | 0.26 | <0.01| <0.01| 0.18 | 0.14 | 0.01 |
| Qtz       | −0.31 | 0.16    | −0.14 | 0.26   | −0.16 | 0.13 | 0.57  | −0.62 | 0.19 | 0.50 | <0.01| 0.01 | 0.95 | 0.08 |
| Sm        | −0.25 | 0.14    | 0.08 | 0.20   | −0.17 | −0.06 | 0.46  | −0.63 | 0.01 | 0.44 | 0.49 | 0.73 | 0.78 | 0.20 |
| Water depth | −0.07 | 0.12    | −0.17 | 0.02   | 0.08 | 0.24 | 0.18  | 0.10 | −0.20 | −0.20 | 0.37 | 0.05 | 0.06 | 0.65 |
| $\delta^{13}C$ | −0.52 | −0.44   | −0.06 | 0.58   | −0.60 | −0.59 | 0.31  | −0.43 | 0.55 | 0.40 | 0.02 | 0.08 | −0.49 | <0.01 |
| $\delta^{18}O$ | −0.79 | −0.34   | −0.21 | 0.79   | −0.75 | −0.38 | 0.71  | −0.70 | 0.54 | 0.63 | 0.47 | 0.35 | −0.13 | 0.78 |

$p$ values in upper right part of table, $r$ values in lower left part. The basis for the correlation of grain types is the relative abundance in the > 125 µm fraction. The correlation to fines is calculated based on the contribution of matrix (< 125 µm) to the total sediment. Only the mineralogy and isotopic composition of the bulk values were used for the correlation. The correlation between isotope data and other parameters are based on a lower number of analyses ($n = 15$) compared to point counting data and mineralogical composition ($n = 45$).

*Correlation matrix: Significant correlations ($p < 0.05$) are marked in bold.*
Fig. 7  Spatial composition map of carbon and oxygen isotopes of sediments on the Panggang reef platform. a $\delta^{13}$C (‰) of 15 bulk sediment samples. b $\delta^{18}$O (‰) of 15 bulk sediment samples. c Satellite image of Panggang reef platform and sampling spots

Table 4  Faunal elements in Kepulauan Seribu (Atmadja 1977; Renema 2008; van der Meij et al. 2009; Park et al. 2010; Ueda 2020) and their mineralogy

| Aragonite     | Red algae (Genus) | Brown algae (Genus) | Green algae (Genus) | Bivalve (Family) | Benthic foraminifers (Genus) |
|---------------|-------------------|---------------------|---------------------|-----------------|------------------------------|
| HMC           | Liagora Galaxaura| Padina              | Halimeda            |                 |                              |
|               |                   |                     |                     |                 | Amphisorus                   |
|               |                   |                     |                     |                 | Borelis                      |
|               |                   |                     |                     |                 | Calcarina                    |
|               |                   |                     |                     |                 | Dendritina                   |
|               |                   |                     |                     |                 | Heterostegina                |
|               |                   |                     |                     |                 | Laevipeneroplis              |
|               |                   |                     |                     |                 | Monalysidium                 |
|               |                   |                     |                     |                 | Operculina                   |
|               |                   |                     |                     |                 | Parasorites                  |
|               |                   |                     |                     |                 | Peneroplis                   |
|               |                   |                     |                     |                 | Sorites                      |
| LMC           |                   |                     |                     |                 | Pectinidae (+ ARA)           |
|               |                   |                     |                     |                 | Amphiostegina                |
|               |                   |                     |                     |                 | Elphidium                    |
|               |                   |                     |                     |                 | Pinnidae (+ ARA)             |
that abrasion rates of corals are much slower compared to dominant HMC components, such as coralline algae and echinoderms (Milliman et al. 1974). Increased HMC content in the fine fraction (Table 2) therefore could be related to preferential breakdown of HMC components. However, it remains unclear in how far these laboratory results can be applied to natural settings. Quantitative data on the composition of different grains size fractions from isolated carbonate platforms of Belize were presented by Gischler and Zingeler (2002). Our calculations based on this data indicate that coral fragments are actually increased in abundance relative to echinoderm and red algae fragments in the fine (< 125 µm) fraction as compared to the larger grain size fraction. This is opposite to what would be expected by the laboratory results on mechanical abrasion (Milliman et al. 1974), possibly indicating that other processes, such as bioerosion, are more important in reef settings. The preferential abrasion of HMC components therefore seems not to be the reason for the increased HMC content in the fine fraction. A known source of mud-sized HMC are red algae growing on sea grass (Land 1970; Perry et al. 2019). Patches of sea grasses are common on the Seribu islands and are associated with reduced grain sizes in the sediment, likely due to their baffling effect (Utami et al. 2018). They also might contribute to the increased HMC contents in the fine fraction. However, a comparison of the mineralogical composition between locations with dense and sparse sea grass cover does not show a significant difference with respect to HMC content (Mann–Whitney U test, p = 0.91). Due to the abundance of sea grass, our dataset does not include samples from locations without sea grass.

Alternatively, micritization in northern Belize increases the HMC content of skeletal grains, which later are preferentially abraded and contribute to carbonate muds. Cementation in micro-bioerosion traces is an important mechanism for micritization (Reid and Macintyre 2000). High Mg calcite cement has been observed, besides other mineral phases, in endolithic microborings in corals (Schroeder 1972). Bioerosion and/or micritization are ubiquitous in reefs affected by terrestrial runoff or upwelling in SE Asia (Tomascik et al. 1997; Wilson 2012; Madden and Wilson 2012) including the Seribu Islands (Fig. 2) (Park et al. 1992, 2010). Both processes are most intense at the outer surface of carbonate grains. The higher surface-to-volume ratio of smaller grains could therefore lead to stronger alteration of its mineralogy compared to larger grains. Micritization and bioerosion could therefore have contributed to the increased HMC content in the fine fraction of Panggang (Table 2).

Small amounts of siliciclastic minerals are found in Kepulauan Seribu (Table 1). Quartz content increases with water depth (Table 3), in the inner lagoon and reef front. Quartz and smectite are strongly concentrated in the fine fraction (Table 2), indicating that quartz grains will settle from suspension in low-energy environments but are easily winnowed by waves and currents in high-energy environments. Smectite occurs only in > 6 m water depth within the inner lagoon (Fig. 4), where it likely settles below the wave base. The fact that siliciclastic minerals are found in almost all samples (Fig. 4) shows that Panggang is influenced by terrestrial runoff (Fig. 1). A possible source for the siliciclastics could be Sumatra, since Kepulauan Seribu is relatively isolated from a direct influence from Java and Kalimantan by the regional current system (Tomascik et al. 1997; Jordan 1998). Partially due to this siliciclastic influence, the depth of abundant coral growth in Kepulauan Seribu (0–18 m, Jordan 1998) is typical for equatorial carbonates in SE Asia, but very shallow compared to other regions (0–100 m) with clearer water and lower nutrient input (Wilson 2008).

**Spatial patterns in stable isotope composition of sediment**

Delineating facies based on their δ13C and δ18O values is not possible due to a lack of a significance difference between carbonate facies (Kruskal–Wallis test, p = 0.17 and p = 0.15, respectively). However, the average stable isotope values of the coral grainstone facies (δ13C = 0.0, δ18O = −4.4) is lower compared to the mollusc wackestone facies (δ13C = 0.4, δ18O = −4.0), suggesting some facies control (Fig. 6). Coral abundance shows a significantly negative correlation to δ13C and δ18O values (Table 3). This results from a strong kinetic control on isotope fractionation in corals (McConnaughey 2003). Coral skeletons therefore show a depletion of heavy isotopes and consequently a strong offset from equilibrium values towards more negative isotope ratios (Fig. 6). Corals exert a strong influence on the bulk isotope composition, since they are the predominant skeletal material and major aragonite producer (Fig. 6). Marine mollusc shells, as the second most important sedimentary component (Table 3), do not show kinetic effects in their isotopic composition (McConnaughey and Gillikin 2008). Their δ18O values are often close to equilibrium values (Grossman and Ku 1986) and therefore more positive compared to corals (Fig. 6) (Aharon 1991). Carbon isotope signatures in molluscs are complex, with near equilibrium values e.g. in giant clams (Aharon 1991) but often several ‰ lighter due to metabolic carbon contribution (McConnaughey and Gillikin 2008). In reefal carbonates, molluscs are typically less depleted in 13C compared to cooccurring corals (Fig. 6) (González and Lohmann 1985). This results in a positive (negative) correlation (Table 3) between oxygen and carbon isotopes and molluscs (corals). Higher stable isotope values in the inner lagoon and sand apron of Panggang (Fig. 7) result from the increased abundance of molluscs in mollusc wackestone and coral packstone/grainstone facies (Table 1) and coincides with the overall lowest aragonite value (Fig. 4). The most
negative δ13C and δ18O values occur at the reef front in coral grainstone and coral packstone/grainstone facies where coral cover is densest (Fig. 7). However, the compositional differences between individual facies types (Table 1) are not strong enough to result in separate isotope fields for each facies (Fig. 6).

The most important group of purely calcitic components are benthic foraminifers. Many shallow water benthic foraminifers from reefal environments precipitate their skeleton close to oxygen isotope equilibrium (± 1‰) (Saraswati et al. 2004; Maeda et al. 2017), although slightly stronger depletion of 18O was observed as well (Wefer and Berger 2004; Maeda et al. 2017), although slightly stronger close to oxygen isotope equilibrium (± 1‰) (Saraswati minifers from reefal environments precipitate their skeleton are benthic foraminifers. Many shallow water benthic foraminifers as the main aragonite and calcite producers, resulting from the strong positive (negative) correlation with molluscs (corals). This is supported by the observation that the fine fraction samples show no consistent isotopic difference to the bulk samples (Table 3). This similarity indicates that the fine fraction is genetically related to the coarse fraction and is formed by skeletal breakdown.

Comparison with other carbonate systems

To facilitate the comparison of mineralogical and stable isotope signatures between humid equatorial, arid (sub) tropical, and temperate carbonate systems we have grouped them geographically (Figs. 8 and 10). Carbonates from each geographical group share common characteristics and are deposited under similar environmental conditions (Fig. 9, Table 5). Carbonate systems from the tropical western Atlantic (TWA) and Persian Gulf represent mostly (sub)tropical carbonates (sensu Wilson 2002, 2012) characterized by more positive δ13C values compared to most calcitic skeletal grains, resulting in a positive correlation between both parameters (Swart et al. 2009; Gischler et al. 2012). Corals on the other hand are more depleted in δ13C compared to most calcite secreting organisms (Fig. 6), which explains the negative correlation between aragonite and δ13C in the coral dominated Kepulauan Seribu (Table 3) and the lower δ13C compared to the relatively calcite-rich cool water carbonates (Fig. 8). However, the δ13C signature of carbonate from Kepulauan Seribu is also more negative compared to most other coral dominated reef facies from the IPWP (Fig. 8).

Due to mineralogical effects, inorganic aragonite is enriched in δ13C by about 1.8‰ compared to coprecipitated calcite (Rubinson and Clayton 1969; Romanek et al. 1992). This mineralogical effect should favor a positive correlation between aragonite content and δ13C. However, the actual relation depends on the source and therefore isotopic signature of the aragonite. Aragonitic non-skeletal grains and green algae (Fig. 8) are characterized by more positive δ13C values compared to most calcitic skeletal grains, resulting in a positive correlation between both parameters (Swart et al. 2009; Gischler et al. 2012). Corals on the other hand are more depleted in δ13C compared to most calcite secreting organisms (Fig. 6), which explains the negative correlation between aragonite and δ13C in the coral dominated Kepulauan Seribu (Table 3) and the lower δ13C compared to the relatively calcite-rich cool water carbonates (Fig. 8).
both of which are located within the IPWP (Fig. 8) (Pfeiffer et al. 2004; Pfeiffer and Dullo 2006; Cahyarini et al. 2016). It is unlikely that this difference is solely due to kinetic related offsets, since all three coral colonies show similar growth rates (Pfeiffer et al. 2004; Pfeiffer and Dullo 2006; Cahyarini et al. 2016). More negative δ13C values of the *Porites* colony from Bidadari Island are more likely due to the influence of river run off. Corals have the potential to record land–ocean carbon transfer because they draw on dissolved inorganic carbon (DIC) for calcification. The δ13C of DIC in coastal water is influenced by the δ13C signature from adjacent rivers (Moyer and Grottoli 2011). Bidadari Island is located in the Java Sea (Fig. 1) and in close proximity to three major rivers in the Greater Jakarta Area (Farhan and Lim 2012). Estuaries in SE Asia show low δ13C values due to the contribution of 12C from riverside mangroves (Miyajima et al. 2009). The low δ13C of the coral skeleton in Kepulauan Seribu therefore likely reflects the enrichment in 12C of ambient seawater due to the input of riverine DIC to the Java Sea.

To a somewhat lesser extent, light intensity (photosynthesis) also has been shown to influence the value of skeletal δ13C in corals (Moyer and Grottoli 2011). Corals are usually depleted in 13C relative to ambient seawater as a result of kinetic and metabolic fractionation (Swart 1983; McConnaughey 1989; Allison et al. 1996; Grottoli and Wellington 1999). Environmental variables that influence coral metabolism should therefore also affect δ13C levels in coral skeletons (Grottoli 2002). Under low light conditions photosynthesis decreases and metabolic fractionation increases with the consequence of decreased skeletal δ13C (Grottoli 2002). Sediment input from rivers into the Java Sea potentially enhances turbidity in the semi-enclosed basin, leading to a reduction of light intensity and photosynthesis thus resulting in reduced skeletal δ13C values. This would be consistent with the relatively shallow depth of reef growth.
Jordan 1998) and the presence of siliciclastic minerals in sediment samples below the wave base (Fig. 4) in the study area. Turbidity can also be produced by algae blooms caused by increased nutrient input either through river water or untreated sewage from populated islands (Baum et al. 2015). Additionally, higher nutrient availability could influence the δ¹³C of coral skeletons by an increase in the heterotrophic feeding rate or an increase in the photosynthesis rate related to higher zooxanthellate concentrations (Grottoli 2002). However, these processes have opposing effect on the δ¹³C of coral skeletons. The net effect therefore will depend on the relative magnitudes of these various influences but is expected to be small (Heikoop et al. 2000).

The largest single anthropogenic influence on the isotope signature of recent corals likely is caused by the input of anthropogenic CO₂ to the global surface ocean. The accelerated admixture of fossil fuel derived CO₂ leads to a shift towards lower δ¹³C values in atmospheric CO₂, the so-called Suess effect (Keeling 1979). This Suess effect is often recorded in declining coral skeleton δ¹³C values over the twentieth century (Swart et al. 2010). Linsley et al. (2019) showed that the Suess effect can shift skeletal δ¹³C in corals by more than 1‰ since the 1950s. The carbon isotope composition of corals in Fig. 8, which are largely derived from the twentieth century, is therefore likely depleted relative to bulk sediments or sedimentary grains of mixed and/or unknown age.

Florida Bay is the only location that shows a nearly complete overlap but broader total range in δ¹³C values compared to the sediments from Kepulauan Seribu (Fig. 8). In contrast to many other areas in the TWA, non-skeletal grains and inorganic mud are lacking in Florida Bay (Table 5). Average δ¹³C of surface waters in Florida Bay is lowest in areas influenced by isotopically depleted river runoff from the Everglades, while the δ¹³C variability in other areas is controlled by remineralization of marine organic matter (Swart and Price 2002). Similar to Kepulauan Seribu, the lowest values in δ¹³C of carbonates deposited in Florida Bay

![Fig. 9 Main controlling factors for oxygen isotope composition of carbonate minerals are temperature and δ¹⁸O of water (a). Annual mean sea surface temperature (World Ocean Atlas 2009; http://www.nodc.noaa.gov/OC5/WOA09/pr_woa09.html). b Global annual mean surface (upper 50 m) δ¹⁸O distribution simulated by a general circulation model (MITgcm) in comparison to the observational isotope data (colored symbols—Schmidt et al. 1999) for the global ocean](image-url)
seem to be a result of isotopically depleted river runoff and a lack of inorganic carbonate precipitation.

Since the pioneering work by Urey (1947), the δ18O values of carbonate minerals are known to be controlled by temperature of formation. Other controlling factors are the δ18O value (δ 18Ow) of the precipitating fluid (Urey 1947; Epstein and Mayeda 1953), mineralogy (Tarutani et al. 1969), and “vital effects” in carbonate secreting organisms (McConnaughey 2003). The slope of the δ 18O/temperature relationship in carbonates is commonly assumed to be in the range from −0.2 to −0.25 ‰/°C (Grossman and Ku 1986; Bemis et al. 1998), resulting in a decrease of 0.8–1 ‰ δ18O for a four degree increase in temperature.

Carbonates from Kepulauan Seribu are characterized by much lower δ 18O values compared to other carbonate systems (Fig. 8). This can be explained by the combined effect of high sea surface temperatures and low δ18Ow in the Java Sea (Fig. 9). Modelling results confirm that for carbonates precipitated in equilibrium with seawater, SE Asia is globally the region with the most negative δ 18O signature (Fig. 9c). Within the IPWP the differences in temperature are small (Fig. 9a), but δ18O water is significantly reduced in SE Asia compared to e.g. the central equatorial Indian Ocean or Pacific (Fig. 9b).

Differences in hydrology are reflected in the lower δ 18O values of a Porites coral from Kepulauan Seribu compared to those from the Seychelles and Chagos (Fig. 8). The difference of more than 1 ‰ in δ18O is more than can be explained by intercolony offsets (Sayani et al. 2019) or the temperature difference of ~1 °C between the Java Sea and

Table 5 Location and basic environmental and sedimentological data for carbonate systems used for comparison in Figs. 8 and 10

| Location                        | SST (°C) | SSS (‰) | δ13C (‰) | δ18O (‰) | Non-skeletal grains | Source                        |
|---------------------------------|----------|----------|-----------|-----------|--------------------|-------------------------------|
| Panggang, Kepulauan Seribu      | 28−30    | 30−34    | 0.02 ± 0.7 | −4.27 ± 0.4 | Very rare          | This study                    |
| Rasdhoo & Ari atoll, Maldives   | 28−30    | 34.4−35.4| 1.56 ± 0.58| −2.06 ± 0.83| Very rare          | Gischler (2006)               |
| Enewetak                        | 28−32    | 34−35    | 1.75 ± 0.55| −1.90 ± 0.42| None               | Weber and Schmalz (1968)      |
| Honduras, Caribbean oceanic     | 29−31    | 36−37    | N/A       | N/A       | Peloids common,     | Schmitt and Gischler (2017)   |
| Barrier reef                     |          |          |           |           | ooids rare         |                               |
| Belize Barrier Reef             | 24−27    | 30−35    | N/A       | N/A       | Rare               | Gischler et al. (2013)        |
| Belize-Yucatan Platform         | 21.7−31.8| 34.2−42.5| N/A       | N/A       | Rare to moderate   | Gischler et al. (2012)        |
| (Glovers, Lighthouse, Turneffe) |          |          |           |           |                   |                               |
| Belize northern shelf           | 25.5−28.5| 30−34    | N/A       | N/A       | Rare               | Weber and Schmalz (1968)      |
| Belize southern shelf           |          |          |           |           |                   |                               |
| Great Bahama Bank               | 18.5−28.5| 36−46    | 4.8 ± 0.3 | 0.4 ± 0.4 | Abundant           | Reijmer et al. (2009); Swart et al. (2009) |
| San Salvador Island, Bahamas    | 24−30    | 35       | 2.2       | −0.9      | Moderate to abundant| Husinec et al. (2019)        |
| Florida Bay                     | 15−40    | 10−40    | N/A       | N/A       | None               | Ginsburg (1956); Swart et al. (1989); Andrews (1991) |
| Kuwait ramp                     | 13−32    | 38.5−42.5| N/A       | N/A       | Abundant           | Gischler and Lomando (2005)   |
| Spencer Gulf, South Australia   | 10−28    | 36−47    | N/A       | N/A       | None               | O’Connell and James (2015)    |
| Laceypede Shelf (LS) and Great  | 35.9−36.4| N/A      | N/A       | N/A       | None               | O’Connell and James (2015)    |
| Australian (GAB), Southern      |          |          |           |           |                   |                               |
| Australia                       |          |          |           |           |                   |                               |
| Wanganui Shelf, New Zealand     | ca. 15   | 34−35    | N/A       | N/A       | None               | Gillespie and Nelson (1997)   |

*N/A = average and standard deviation data not available*
the other two sites. Most of the difference therefore can be attributed to the contrasting δ¹⁸O seawater values, which based on the difference in salinity should be around 0.9–1.1 ‰/δ¹⁸O assuming a slope of 0.22 ‰ for the salinity/δ¹⁸O relationship (Fairbanks et al. 1997). The lower δ¹⁸O seawater values in the Java Sea result from heavy precipitation and increased runoff, and are also influenced by the δ¹⁸Ow of the precipitation itself via the so called “amount effect” (Rozanski et al. 1993; Lau and Yang 2003).

Facies also could influence the δ¹⁸O signature of bulk sediments due to differences in the “vital effects” of carbonate producing organisms (McConnaughey 2003; McConnaughey and Gillikin 2008) and mineralogy (Tarutani et al. 1969). However, the different δ¹⁸O values for individual grain types to a large extent seem to average out for bulk samples (Gischler et al. 2012). This results in relatively large overlaps between facies in most atolls, barrier reefs and platforms (Gischler et al. 2012), showing that the influence of facies in these settings is limited.

The carbonate mineralogy of shallow water (< 20 m) bulk sediments in the tropics and subtropics (IPWP, TWA) is predominantly composed of aragonite and HMC, with LMC usually present in low quantities < 10% (Fig. 10) (Ginsburg 1956; Purdy 1963; Weber and Schmalz 1968; Gischler 2006; Reijmer et al. 2009), San Salvador, Bahamas (Husinec et al. 2019), Florida Bay (Ginsburg 1956), Kuwait ramp (Gischler and Lomando 2005), Spencer Gulf, South Australia (O’Connell and James 2015), Lacepede Shelf and Great Australian Bight, South Australia (James et al. 2005) are shown for comparison. All sample are from shallow water (< 20 m) except sample from South Australia.

![Mineralogical composition of bulk surface sediment from Panggang reef platform, Kepulauan Seribu (normalized to only carbonate mineral phases). Carbonate mineralogy of Holocene bulk surface sediment from the Maldives (Gischler 2006), Eniwetok Atoll (Weber and Schmalz 1968), Honduras, Caribbean oceanic barrier reef (Schmitt and Gischler 2017), Belize (Weber and Schmalz 1968; Gischler et al. 2012, 2013), Great Bahama Bank (Purdy 1963; Reijmer et al. 2009), San Salvador, Bahamas (Husinec et al. 2019), Florida Bay (Ginsburg 1956), Kuwait ramp (Gischler and Lomando 2005), Spencer Gulf, South Australia (O’Connell and James 2015), Lacepede Shelf and Great Australian Bight, South Australia (James et al. 2005) are shown for comparison. All sample are from shallow water (< 20 m) except sample from South Australia.](https://example.com/figure10.png)
Opposed to open ocean settings, the semi-enclosed Java Sea is influenced by river runoff. Terrestrial sediments reaching Kepulauan Seribu (Table 1) most likely are derived from Sumatran rivers, since oceanic currents isolate Kepulauan Seribu from sediment input from Java and Kalimantan (Tomascik et al. 1997; Jordan 1998). The suspended particulate matter promotes oligophotic conditions in relatively shallow water. Wilson and Vecsei (2005) argued that oligophotic conditions in water depth below 20 m are typical for humid equatorial carbonate systems in SE Asia and would favor the deposition of e.g. relatively calcite-rich foramol facies. Foramol facies in Kepulauan Seribu dominates in water depth below 27 m (Jordan 1998).

In water depth < 20 m, rotaliid foraminifers contribute to LMC (Table 4) but their abundance is not high enough to explain the generally increased LMC content. Another important source of LMC in Kepulauan Seribu seems to be mixed aragonite-calcite bivalve shells. More mineralogical analysis from humid equatorial settings are needed to test if the increased LMC content observed in Kepulauan Seribu is a common feature of shallow water (< 20 m) reef systems in humid, equatorial settings.

Fine fraction sediment from the Panggang reef platform is showing a higher HMC compared to the bulk fraction, likely related to micritization and bioerosion. Higher HMC contents in the fine fraction are also found in Belize, the potential source is micritization of skeletal grains to Mg calcite (Weber and Schmalz 1968). Micritization of skeletal grains is also an important diagenetic process in other shallow marine carbonate platforms, e.g. The Bahamas where the grains are altered to aragonite (Purdy 1963; Bathurst 1966; Kendall and Skipwith 1969; Trumbull 1988). Preferential alteration of the carbonate grains into high Mg calcite is potentially controlled by lower aragonite supersaturation in water off Kepulauan Seribu. Fresh water influx from rivers promotes low salinities in the semi-enclosed seas of SE Asia, leading to relatively low aragonite supersaturation compared to more arid regions as the tropical, western Atlantic (Park et al. 2010). Seawater Mg:Ca ratios are another important control on carbonate mineralogy, with lower Mg:Ca favouring calcite precipitation at a given temperature (Morse et al. 1997). A recent compilation of Mg:Ca ratios in seawater shows that values are not spatially constant but influenced by river waters with low Mg:Ca values (Lebrato et al. 2020). No seawater data is available for the semi-enclosed seas of SE Asia, but the high riverine input could potentially also reduce the Mg:Ca ratios in the Java Sea. Depending on the magnitude of this effect, the reduced Mg:Ca ratios could facilitate the precipitation of calcite in the study area.

The mineralogical composition in the shallow water (< 20 m), photozoan-dominated carbonates from Kepulauan Seribu overlaps with the most aragonite rich part of the cool water carbonate field (Fig. 10). However, based on the dominance of reefal material, the sediments can be clearly differentiated from heterozoan dominated cool water carbonates. This might be more difficult for the foramol facies that dominates at the toe of slope and below (Jordan 1998). Wilson (2012) pointed out that oligophotic facies that are preferentially deposited in deeper waters below the wave dominance of reefal material, the sediments can be clearly differentiated from heterozoan dominated cool water carbonates. This might be more difficult for the foramol facies that dominates at the toe of slope and below (Jordan 1998). Wilson (2012) pointed out that oligophotic facies that are preferentially deposited in deeper waters below the wave base, is the widespread presence of > 10% LMC content in the shallow water sediment. The LMC is contributed by river waters with low Mg:Ca values (Lebrato et al. 2020).
by organisms that are not or less light dependent (bivalves and some rotaliid foraminifers) and therefore can thrive under oligophotic conditions. The suspended particulate matter from terrestrial runoff likely limits the depth of the euphotic zone. This is reflected in the deposition of foramin facies below a depth of ~28 m (Jordan 1998), but seems to already affect the sediments in the upper 20 m. Compared to similar foramin facies from higher latitudes, the Mg content in HMC components from Kepulauan Seribu is much higher due to the different water temperature between both regions.

The $\delta^{13}C$ signature of carbonates from Kepulauan Seribu is considerably more negative compared to nearly all other platform and reef systems. This is likely due to the combination of low $\delta^{13}C_{\text{DIC}}$ in surface waters and the lack of green algal grains and inorganically precipitated carbonates. The $\delta^{13}C$ in surface waters likely is reduced by $^{12}C$-enriched input of DIC to river water from riverside mangroves. Corals as the predominant skeletal material and major aragonite producer also exert an influence on the bulk isotope composition, resulting in depletion of heavy isotopes values due to kinetic “vital effects”. Low light intensity due to turbid water in the semi-enclosed basin of the Java Sea potentially hampers photosynthesis, which further reduces skeletal $\delta^{13}C$ values in corals due to metabolic effects. The Suess effect, which is caused by the combustion of fossil fuel, leads to a decline of skeletal $\delta^{13}C$ over the twentieth century in corals. This signal might not be recorded to its full extent in the bulk sediment yet, which represents a grain mixture of different ages.

Carbonates from Kepulauan Seribu are also characterized by much lower $\delta^{18}O$ values compared to other carbonate systems, due to high sea surface temperatures and low $\delta^{18}O_{\text{w}}$ in the Java Sea. Deep atmospheric convection leads to heavy precipitation and increased fluvial runoff of very $^{18}O$-depleted water to the semi-enclosed Java Sea. The relatively low $\delta^{18}O$ of benthic foraminifers and bivalves from Kepulauan Seribu shows that the oxygen isotope signature can be used to distinguish foramin facies from oligophotic equatorial carbonates and cool water carbonates in the rock record.

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