Hardened faecal pellets as a significant component in deep water, subtropical marine environments

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Abstract
Non-skeletal carbonate grains are classically interpreted to form in shallow, tropical environments. Peloids deposited in deep, subtropical marine conditions are poorly studied. IODP site U1460 on the subtropical Carnarvon Ramp (Southwest Shelf of Australia) recovered a nearly continuous Pliocene to Recent record of outer shelf and slope sediments. The relative abundance of peloids varies between 0% and 67% of the fine to medium sand fraction, and contributes on average ~4% of all grains. The origin and composition of these peloids were investigated using scanning electron microscopy equipped with an energy-dispersive X-ray spectrometer, light microscopy, X-ray diffraction and stable isotope analysis. The peloids have a uniform size and shape and are interpreted as faecal pellets. They are mainly composed of skeletal fragments such as ascidian spicules, planktic foraminifera and sponge spicules in a mud-sized matrix containing abundant coccolith plates. Mineralogical analysis show that the pellets consist of aragonite, calcite and dolomite. The pellets have an identical mineralogical composition and skeletal assemblage as the surrounding matrix, indicating that they have formed in situ. They occur more abundantly during interglacials when the site was situated in deeper waters below the swell wave base, presumably because the pellets were protected from disintegration and therefore available for cementation. The presence of framboidal pyrite within the pellets indicates bacterial sulphate reduction (BSR). The reduction of iron by hydrogen sulphide produced during BSR decreases the pH and likely explains the observed aragonite dissolution. Aragonite dissolution likely increases the alkalinity, and in consequence causes the precipitation of calcite and dolomite cements. It is suggested here that pellets are hardened due to this early cementation close to the sea floor increasing the potential for preservation in the fossil record.

KEYWORDS
Australian Shelf, early cementation, IODP, non-skeletal grain, peloid, subtropical carbonate

1 INTRODUCTION

Non-skeletal grains are typical components of many shallow water, tropical carbonate environments in the Atlantic Ocean and Caribbean Sea, e.g. the Belize-Yucatan system (Gischler & Lomando, 1999; Milliman, 1969, 1974), the Florida shelf (Ginsburg, 1956) and the Bahama Banks (Bathurst, 1975; Enos, 1974; Harris, 1979; Imbrie & Purdy, 1962; Neumann & Land,
Generally, peloids occur less abundantly in the Indo-Pacific region but are still present, e.g. in the shallow water lagoon of Bora Bora, in areas of elevated alkalinity and pH (Gischler, 2011). Ooids and peloids on the Northwest Shelf of Australia formed in shallow water under arid conditions during the mid to late Pleistocene (Christensen et al., 2017; Gallagher et al., 2018; James, Bone, Kyser, Dix, & Collins, 2004). In contrast to tropical settings, peloids formed as faecal pellets are the only non-skeletal carbonate grains which occur in subtropical to cool-water carbonate settings. However, under the influence of waves and currents, these faecal pellets typically disintegrate into mud-sized particles and are therefore not preserved as grains (Farrow & Fyfe, 1988). Consequently, hardened faecal pellets were described only as a very minor component from cool-water carbonate settings, such as the Great Australian Bight and Spencer Gulf (Betzler, Saxena, Swart, Isern, & James, 2005; O’Connell & James, 2015). Peloids have been recently reported from IODP site U1460 on the subtropical Southwest Shelf of Australia (Gallagher et al., 2017, 2018). The aim of this study was to analyse the origin of these peloids using modal analysis and scanning electron microscopy (SEM) equipped with an energy-dispersive X-ray spectrometer (SEM-EDX). The depositional and diagenetic conditions under which peloids form and are preserved will be determined for subtropical and cool-water carbonates, with important implications for their interpretation in the fossil record.

2 | GEOLOGICAL SETTINGS

The environment of the south-western continental margin of Australia in the eastern Indian Ocean is transitional between warm-temperate and tropical carbonate settings. It comprises the warm-temperate Rottnest Shelf in the south (south of 28°S) and the subtropical Carnarvon Ramp in the north (22–28°S). The Houtman Abrolhos Reef complex, which contains the southernmost major tropical reef in the Indian Ocean (28–29.5°S), straddles the boundary between the Carnarvon Ramp and the Rottnest Shelf (James, Collins, Bone, & Hallock, 1999; Figure 1). The modern-shelf sediments are characterized by a distinct cool-water composition (Nelson, 1988), but

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**FIGURE 1** (a) Location map of IODP site U1460, Tamala line (N, red circles) and sedimentary facies on the modern sea floor of the Carnarvon Ramp/northern Rottnest Shelf, along Southwest Shelf of Australia. Bathymetric contours are in metres (modified after James et al., 1999). (b) Cross-section of Tamala line (N) with distribution of modern facies and location of the IODP site U1460 (modified after James et al., 1999)
with subtropical attributes (James, Bone, Hageman, Feary, & Gostin, 1997). The skeletal assemblage is dominated by coralline algae, bryozoans, molluscs (scaphopods, bivalves and gastropods) and foraminifers (James et al., 1997). The main difference from typical cool-water deposits is the presence of scattered zooxanthellate corals and large, symbiont-bearing foraminifers (Collins, James, & Bone, 2014; James et al., 1999). Skeletons of serpulid worms, echinoids, azooxanthellate corals and sponge spicules are of local importance. The calcareous green algae 
*Halimeda* grows locally on the shelf and ramp but is poorly calcified and therefore generally does not contribute to the sediment outside Shark Bay. The sea floor above ~50–60 m is subject to constant reworking and abrasion by waves, while the swell wave base is close to 100 m, leading to a lack of mud deposition above this depth.

The Carnarvon Ramp comprises the Ningaloo Reef and hypersaline Shark Bay on the inner ramp. The mid-ramp is euphotic, with relatively little calcareous benthos and with low numbers of bryozoans, coralline algae and larger foraminifers, but is dominated by relict or stranded foraminiferal-dominated sand. The outer ramp is pelagic in character, covered by planktic foraminiferal sand or spiculitic mud. The depression of temperature and salinity contours indicates the influence of periodic downwelling and the outflow of highly saline waters from Shark Bay (Collins et al., 2014; James et al., 1999). The Rottnest Shelf is flat-topped, with a wave-swept inner shelf plain characterized by rhodolith pavements, bryozoans, sponges and abraded sediments. An incipient rim formed by a linear ridge system covered by rhodolite gravel separates the inner shelf from the euphotic outer shelf that is dominated by bryozoans, benthic foraminifers and molluscs. The upper slope contains fine sand and silt of bryozoan fragments, sponge spicules and planktic foraminifers (James et al., 1999).

IODP site U1460 was drilled during IODP Expedition 356 on the Southwest Shelf of Australia in a water depth of 214.4 m at 27°22.4867'S and 112°55.4265'E (Figure 1). It is situated north of the Houtman Abrolhos reef complex, at the transition between the Carnarvon Ramp in the north and the Rottnest Shelf towards the south (Figure 1a,b). This transition zone shows a mixture of morphologic and facies characteristics of both regions. James et al. (1999) described the shelf morphology and facies distribution on the modern sea floor. The facies on the inner shelf is characterized by skeletal sands (facies A1; Figure 1) and gravel, with the siliciclastic content increasing towards the coast (facies A2; Figure 1). A high abundance of intraclasts and other relict and stranded grains characterizes the nearshore zone down to a water depth of ~50 m (facies I1; Figure 1). The ridge complex in the transition zone is deeper and morphologically less well defined compared to the Northern Rottnest Shelf. It separates the relatively deep inner shelf from a more gently inclined outer ramp/shelf. The ridge complex, which is interpreted as being a late Pleistocene beach-dune complex (James et al., 1999), is covered by rhodolite gravel and subordinate skeletal sand (facies A3; Figure 1). In the transition zone, the sediment on the ridge complex is relatively rich in bryozoans. The outer ramp/shelf extends from the base of the ridge complex to the shelf edge at approximately 200 m water depth (mwd). The dominant facies on the shallower part of the outer ramp/shelf is bryozoan skeletal sand (facies B; Figure 1), while the deeper part is characterized by pelagic sand facies (P1, P2). The pelagic sand facies is situated below the swell wave base and therefore is the shallowest facies containing carbonate mud (>20%). The content of azooxanthellate corals in this facies is relatively high, with a ~10–20% contribution to the coarse fraction (James et al., 1999). The continental slope in more than 200 mwd is dominated by the carbonate silt facies (S), containing ~70% mud, 30% sand and only trace amounts of coarse fraction (>2 mm). The sand fraction is composed mainly of 95% carbonate with trace terrigenous grains. Planktic foraminifers and sponge spicules contribute most skeletal elements to the sand fraction, while bryozoans, ostracods, pteropods and benthic foraminifers are also common. Worm tubes and faecal pellets are further sediment constituents that can be locally abundant. Azooxanthellate corals, in contrast, seem to be absent (James et al., 1999).

### 3 | MATERIAL AND METHODS

From Hole U1460B, a 306 m long hydraulic piston core with 98% recovery was obtained (Figure 2). The IODP site U1460 consists of two lithostratigraphic units (I, II), with Unit I subdivided into three subunits (Ia, Ib, Ic) (Gallagher et al., 2017). In this study, the focus is placed on the interval between 0 and 108 m core depth below sea floor (CSF-A) belonging to Subunits Ia (0-44.94 m CSF-A) and Ib (44.94-173.36 m CSF-A) (Figure 2).

Unit I as defined in Gallagher et al. (2017) consists mainly of unleithified to partially lithified skeletal packstone with minor skeletal wackestone, and some lithified intervals (Figure 2). This unit is divided into three Subunits based on the abundances of macrofossils and sponge spicules and variation in diagenetic alteration. Macrofossils are concentrated in Subunit Ia, and sponge spicules in Subunit Ib (Gallagher et al., 2017). Subunit Ia is characterized by beige to greenish-grey skeletal packstones. The packstones are interbedded with skeletal wackstones and mudstones in the uppermost 20 m (CSF-A). The sediment is predominantly unleithified, except for a minor lithified interval in the lowermost part of Subunit Ia. Bioclasts of neritic macrofossils such as echinoderms, bryozoans, bivalves and gastropods are present, with major pelagic components including planktic foraminifers. The base of the lithified interval in Subunit Ia is defined by a hardground. The upper part of Subunit Ib is dominated by
unlithified beige to light brown skeletal packstone and the lower part by partially lithified skeletal wackestone and mudstone (Gallagher et al., 2017). The cores U1460-1F to 25F (0–108.6 m CSF-A) were sampled with a 10 cm³ tube every 20 cm on average. This interval represents a time from 0 to 762 ka, following the age model of Petrick et al. (2018), who tuned biomarker derived Sea Surface Temperature estimates to the benthic isotope stack LR04 at this site. This age model is also consistent with the last occurrence of the planktic foraminifer *Globorotalia tosaensis* (0.62 Ma) at 86.5 m CSF-A (Gallagher et al., 2017). Marine isotope stages (MIS) were assigned according to Petrick et al. (2018).

All 103 samples were wet sieved through a 63 μm sieve, and the retained coarser fraction was subsequently dry sieved through 125, 500 and 2,000 μm sieves. The mud fraction (<63 μm) of a subset of 24 samples was further wet sieved through a 34 μm sieve. All individual grain-size fractions of these samples were weighed. A small portion of the bulk sediment from the same subset was obtained for mineralogical analysis using X-ray diffraction (XRD). Samples were oven-dried, ground and mounted on sample holders. The measurements were conducted using a Siemens D5000 X-ray diffractometer over an angle field of 60° (4°–64°) with a step size of 0.02° per second. Identification and quantification of different mineral phases was achieved with the software DIFFRAC EVA (ver. 8.0) by Bruker. The relative abundance of mineralogy of peloids was achieved by 2-D XRD measurements (Smidej et al., 2015).

The abundance of peloids in the 125–500 μm fraction of all 103 samples was quantified with a point counter by the point counter.
counting 300 grains per sample. The length and width of 89 peloids from a total of nine samples was measured under the binocular microscope. A total of 16 thin sections from the first 70 m (CSF-A) were prepared for petrographic analysis of the sand and gravel fraction.

Scanning electron microscopy was used to analyse the composition of pellets and the grain fractions <34 μm and 34–63 μm. The mineralogy of grains was confirmed by elemental analysis (Sr, S, Fe, Mg) with an energy-dispersive X-ray spectrometer (EDS). All samples were carbon coated prior to analysis.

Ten peloid samples from the upper 40 m (CSF-A) were selected for stable isotopic analyses (δ¹⁸O and δ¹³C), which were performed at Leipzig University. Carbonate powders were reacted with 105% phosphoric acid at 70°C using a Kiel IV online carbonate preparation line connected to a MAT 253 mass spectrometer. Isotope values were calibrated to the Vienna Pee Dee Belemnite (V-PDB) standard using the NBS-19 carbonate standard. Average standard deviation associated with the analyses of a reference standard is <0.050‰ V-PDB or δ¹⁸O and <0.016‰ V-PDB for δ¹³C.

4 | RESULTS

4.1 | Composition, mineralogy and texture of sediment

The average carbonate content of the sediment is ~90%, with 57% low-magnesium calcite (LMC), 12% aragonite, 12% high-magnesium calcite (HMC) and 9% dolomite (Table 1; Figure 2e). The aragonite content of the bulk sediment decreases with depth from around 30% to 4% at the base of the section (Figure 2f). Above ~50 m (CSF-A), the mean HMC content is 26%, but HMC disappears completely below this depth. At around the same depth, the dolomite content increases slightly and reaches a maximum value of 21% at a depth of 94.5 m CSF-A in the lower half of the studied interval. The siliciclastic fraction consists mainly of quartz and plagioclase feldspar (Figure 2c). Sulphate minerals such as celestite and anhydrite occur locally as a minor component.

The bulk sediment of the whole interval consists of 3% gravel (>2 mm), 54% sand (2 mm–63 μm) and 43% mud (<63 μm) (Table 1). Generally, the mud content increases with depth (Figure 2d). The gravel consists mainly of slightly cemented lumps formed after burial and larger sporadic bioclasts.

The fossil assemblage of the sand fraction is largely composed of benthic and planktic foraminiferal tests, bivalves, echinoids, bryrozoans, pteropods, serpulids and some solitary corals. The coarser mud fraction (63–34 μm) consists predominantly of skeletal fragments of benthic and planktic foraminiferal tests, ascidian spicules, mollusc shells, echinoderms and sponge spicules (Figures 3b and 4a). Cocolith
plates and ascidian spicule rays are most abundant in the finer mud fraction (<34 μm; Figure 4b).

Subunits Ia and Ib were defined onboard the Joides Resolution based on differences in skeletal assemblage, mineralogy and diagenetic alteration (Gallagher et al., 2017). This subdivision is confirmed by the data presented here. Subunit Ia is characterized by a relatively high content of aragonite (19%, Table 1) and HMC (25%, Table 1). Based on visual estimates, the skeletal assemblage is dominated by planktic foraminifers, mollusc fragments (including pteropods), bryozoans, echinoids and some azooxanthellate corals, intraclasts and peloids. The sediment is a mixture of sand (64%, Table 1) and mud (33%, Table 1). The gravel content is very low (3%, Table 1), except for some coarse-grained intercalations that consist of sand (63%) and gravel (37%). The sand fraction in these layers consists predominantly of bryozoan fragments, the rest comprising variable proportions of foraminifers, molluscs, serpulids, echinoids and minor amounts of azooxanthellate corals, ascidians spicules and scaphopods. The gravel fraction has a similar composition but contains many skeletal intraclasts and is lacking ascidian spicules. Peloids and quartz grains are generally absent in all of these coarse-grained intercalations.

The investigated part of Subunit Ib is characterized by a lower aragonite (~7%, Table 1) content compared to Subunit Ia and by a near absence of HMC as well as an increased dolomite content (~12%, Table 1). The skeletal assemblage is similar to Subunit Ia, but the number of macrofossil fragments is generally lower. Further differences are the high content of sponge spicules, which were much less in subunit Ia, and the absence of azooxanthellate corals (Figure 2). The content of peloids at Subunit Ib increases compared to Subunit Ia (14%, Table 1) and intraclasts are absent from the sand fraction. Overall, Subunit Ib has a lower sand (47%, Table 1) and a higher mud content (50%, Table 1) compared to subunit Ia. Gravel contributes about 3% (Table 1) to the sediment.

### 4.2 | Morphology and composition of peloids

The ovoid to ellipsoid shaped peloids are generally medium to coarse sand-sized, with a mean length of ~650 μm (SD: 90 μm), a mean width of ~390 μm (SD: 60 μm) and a length to width ratio of ~1.7. Because of their width, the peloids are found nearly exclusively in the 125–500 μm fraction. Over the entire interval, peloids contribute a mean of ~10% grains (range: from 0% to 67%; Figure 2g) to this size fraction (Table 1), or 4% of all grains. Their content nearly doubles from Subunit Ia (~7%, Table 1) to Subunit Ib (~14%, Table 1). Peloid abundance is always higher in each interglacial stage (odd MIS) compared to the overlying glacial stage (even MIS; Figure 2, Table 2). This relationship is most clearly visible in the upper ~30 m CSF-A.

The peloids have a smooth surface, very uniform morphology and size (Figure 3) which characterize them as faecal pellets (Blom & Alsop, 1988; O’Connell & James, 2015). They are mainly composed of skeletal fragments such as ascidian spicules, planktic and benthic foraminiferal tests, and sponge spicules in a micritic matrix containing coccolith plates (Figure 4b–f). Non-carbonate minerals such as pyrite and plagioclase feldspar occur as further locally important constituents. 2D-XRD shows that the pellets have a very similar mineralogical composition to the bulk sediment at the same depth.

### 4.3 | Diagenetic processes and stable isotopes

Aragonite dissolution in the mud-sized matrix and within the faecal pellets is visible at the tips of ascidian spicules (Figures 4c and 5a). Authigenic phases formed within the matrix sediment and the faecal pellets are frambooidal pyrite, HMC (7–9 mol%) and dolomite cement (Figures 5b–d and 6a–e). Phosphate was observed lining the outer rim of faecal pellets (Figure 6f).
The δ13C values of faecal pellets range from 1.2‰ to 2.3‰ with a mean of 1.6‰, and their δ18O values range from 1.19‰ to 1.6‰ with a mean of 1.5‰ (Figure 7).

5 | INTERPRETATION

5.1 | Origin of peloids

At present, IODP site U1460 is situated in a water depth of ~214 m, at the transition between the planktic sand and silt facies (P2) and the carbonate silt facies (S1). A comparison with the composition of present sea floor sediments (James et al., 1999) indicates that the sediments in Subunit 1a are very similar to the planktic sand and silt facies (P2; James et al., 1999), i.e. consisting of abundant planktic foraminifers, relatively few intraclasts and ~33% mud content. On the modern day sea floor, this facies is restricted to the outer ramp and occurs in water depths of between ~120 and 200 m (Figure 1). The majority of the sediment in subunit Ib is therefore interpreted to have formed over a similar depth range. The coarse-grained, gravel-rich sediments, in contrast, are similar in composition to the bryozoan skeletal sand facies (B; James et al., 1999) on the modern day sea floor. The main similarities to this facies are the amount of gravel, the lack of mud, the absence of quartz grains and the high proportion of bryozoan debris compared to coralline algae fragments. The bryozoan skeletal sand facies is typically deposited on the outer shelf and occurs in water depths of between 100 and 150 m, landward of Site 1460 (Figure 1, James et al., 1999). Based on their correlation with glacial MIS (Figure 8), at least some of the gravel-rich layers were likely deposited in situ during glacial sea-level lowstands and therefore do not represent drilling disturbances (fall-in deposits) as interpreted by Gallagher et al. (2017) in the shipboard reports.

The investigated part of Subunit Ib shows many similarities with the spiculitic carbonate silt facies (S1) such as the high mud content, the abundance of sponge spicules and the
absence of azooxanthellate corals. The spiculitic carbonate silt facies is also the only facies from which peloids are described on the modern sea floor (James et al., 1999). Today, this facies occurs in water >200 m deep on the slope of the Carnarvon ramp and the northern Rottnest Shelf (Figure 1). Therefore, Subunit Ib was likely deposited in deeper water compared to Subunit Ia. This is also in accordance with the lower aragonite content in Subunit Ib (Figure 2), as there is a tendency for decreasing aragonite content in modern cool-water environments from the middle and outer shelf towards the upper slope (James, Bone, & Kyser, 2005). Additionally, the decrease in aragonite with depth could be caused by subsea floor aragonite dissolution (James et al., 2005).

Sediments of Subunit I were deposited during several glacial–interglacial cycles, which are expected to influence the skeletal assemblage in the inner and mid-ramp/shelf environments. However, the climate-related changes in skeletal assemblages in the outer ramp/shelf and slope sediments deposited at IODP site U1460 are expected to be subtle since bottom water temperatures at the site were likely always <15°C (James et al., 1999).

The similarity of the skeletal assemblage in both the matrix and the pellets indicates that pellets have formed in situ and were not imported from shallower environments. The abundance of skeletal elements from benthic organisms such as sponge and ascidian spicules indicates that they were formed by deposit-feeders and not by pelagic organisms. The fact that all pellets are very similar in composition, size and shape point to a single group of producers. The ovoid to ellipsoidal shape with a length to width ratio of 1.7:1 is typical for faecal pellets of polychaete worms or molluscs (Bandel, 1974; Martens, 1978). Deposit feeding polychaete worms occur abundantly on the Carnarvon ramp (Brooke et al., 2009) and even form the most prominent group of benthic organisms on soft, fine grained sediments on the Northwest Shelf of Australia (Jones et al., 2007). It therefore seems to be likely that polychaete worms like are the producers of the faecal pellets. Their higher abundance of pellets in interglacial intervals indicates that more pellets were deposited when IODP site U1460 was situated in deeper waters. Additionally, the presented facies interpretation indicates that pellet abundance is higher in the deeper water slope facies (Subunit Ib, 14% of 125–500 μm fraction) than in the more proximal outer ramp/shelf facies (Subunit Ia, 7% of 125–500 μm fraction).

### Table 2

| MIS | N  | Mean peloid abundance (%) | SD   |
|-----|----|---------------------------|------|
| 1   | 4  | 0.3                       | 0.7  |
| 2–4 | 5  | 0                         | 0.0  |
| 5   | 4  | 3.3                       | 1.8  |
| 6   | 3  | 0.2                       | 0.4  |
| 7   | 3  | 2.4                       | 4.2  |
| 8   | 4  | 0.4                       | 0.7  |
| 9   | 2  | 5.3                       | 2.4  |
| 10  | 2  | 0                         | 0.0  |
| 11  | 11 | 16.6                      | 9.7  |
| 12  | 11 | 5.2                       | 5.2  |
| 13  | 8  | 14.1                      | 9.4  |
| 14  | 4  | 14.4                      | 13.0 |
| 15  | 34 | 14.7                      | 15.4 |
| 16  | 2  | 1.5                       | 0.7  |
| 17  | 2  | 5.7                       | 5.2  |

FIGURE 5 Cementation and aragonite dissolution in the mud matrix (grain size fraction <63 μm). A: Voids at the tips of ascidian spicule (a) rays indicate aragonite dissolution (white arrows; 356-U1460B-12F-2-W 68/72). B: Framboidal pyrite (p) formed during BSR. C: High-Mg calcite cementation (c: 356-U1460B-5F-2-W 79/83). D: Dolomite cementation (d: 356-U1460B-9F-2-W 14/18)
5.2 Aragonite dissolution

Bioclastic aragonite at IODP site U1460 was produced mainly from the shells and skeletal elements of ascidians, pteropods and bivalve shells via maceration. Maceration is the breakdown of grains along lines of weakness between skeletal subunits into their microscopic structural elements, needles and granules (Figure 6c; Alexandersson, 1979; O’Connell & James, 2015). The observed systematic decrease in aragonite content with depth could be due to platform progradation (see above) and dissolution during shallow burial (Figure 2).

The presence of the frambooidal pyrite in pellets already at their shallowest occurrence (1.15 m CSF-A) indicates active bacterial sulphate reduction (BSR) near the sea floor.

5.3 Stable isotope signature of peloids

The studied faecal pellets from the outer shelf of the subtropical Carnarvon ramp show higher oxygen and lower carbon isotopes values compared to their tropical counterparts, e.g. from the modern arid carbonate ramp of Kuwait or the isolated carbonate platforms of the Belize-Yucatan system (Gischler & Lomando, 2005; Gischler, Swart, & Lomando, 2009) (Figure 7). In contrast, the values are very similar to isotope values of sea floor carbonate sediments and HMC cements from the cool-water, southern Australian Shelf (Figure 7). The isotopic signature therefore might be a useful discriminator for pellets from these two environments.

6 DISCUSSION

6.1 Origin of peloids

Non-skeletal grains are typically formed in shallow, tropical environments such as the Bahamas (Bathurst, 1975; Harris, 1979; Hine, 1977; Lloyd, Perkins, & Kerr, 1987; Purdy, 1963). Sediment sampling at the sea floor indicated the presence of faecal pellets on the slope of the Southwest Shelf of Australia at a depth >200–300 m (James et al., 1999). Several previous studies on temperate to cool-water regions indicate that faecal pellets disintegrate easily into mud-sized carbonates (Farrow & Pyfe, 1988) and that hardened faecal pellets occur only sporadically in temperate, shallow marine carbonates (O’Connell & James, 2015). This is consistent with the lack of faecal pellets in waters shallower than 200 m on the swell-dominated Southwest Shelf of Australia (James et al., 1999). However, Blom and Alsop (1988) have demonstrated that faecal pellets accumulate in relatively shallow water at depths of between 70 and 85 m in the central Bass Basin (Southeast Australian Shelf), where they are protected...
from swells. The data presented here show that faecal pellets form a significant part of the sand fraction of the middle to upper Pleistocene interval at IODP site U1460. The faecal pellet abundance shows a clear covariance with relative water depth since pellets occur preferentially in the slope facies (Subunit Ib) and during interglacial intervals when IODP site U1460 was situated in relatively deep water. The changes between glacial and interglacial intervals are more clearly expressed in the upper ~30 m CSF-A (MIS 1-12), which was influenced by high-amplitude sea-level variations. For example, during the Last Glacial Maximum, when sea-level dropped by ~120 m (Lisiecki & Raymo, 2005), the water depth at IODP site U1460 would have been around ~90 m. This would have brought the site above the present day swell wave base (James et al., 1999), causing pellets to disintegrate (Figure 8b; glacial MIS1-12). Compared to previous glacial–interglacial cycles, MIS 13-15 were possibly characterized by relatively lower amplitude sea-level fluctuations (Miller et al., 2005), insufficient to place the site above the swell wave base and potentially explaining the lack of a clear correlation between MIS and pellet abundance in this interval. A more important reason might be that the depositional environment was generally deeper in the lower part of the interval, as indicated by a shift to slope facies in Subunit Ib, dampening the effect of sea-level fluctuations.

6.2 Comparison to peloids from tropical environments

This study shows that faecal pellets can be formed and preserved in temperate carbonates on the transition from the outer ramp/shelf to slope, where they are protected below the swell wave base from disintegration. Volumetrically, they form up to 67% of the fine to medium sand fraction and on average ~2% of the total sediment of the entire interval. The abundance of pellets in these relatively deep-water temperate carbonates is therefore similar or higher than most tropical carbonate shoals and shallow water platforms in the Indo-Pacific region (Gischler, 2011; Utami, Reuning, & Cayharinin, 2018). Many carbonate platforms and shoals in the Caribbean (Milliman, 1967, 1969) have higher peloid contents compared to IODP site U1460, but several have similar or even lower peloid abundances, for example the Dry Tortugas shoal on the distally steepend ramp of the west Florida shelf or the Turneffe Islands platform off Belize.
(Gischler, Isaack, & Hudson, 2017; Gischler & Zingeler, 2002). The skeletal assemblage contained in those shallow water, tropical faecal pellets is partially similar to the pellets from deeper water, temperate carbonates described in this study. Gischler (2011) mentioned that cemented faecal pellets from Bora Bora contain skeletal components such as mollusc and (benthic) foraminiferal shell fragments, sponge needles or tunicate spicules. However, the tropical shallow water pellets lack the pelagic component that is common in the pellets at IODP site U1460.

Another difference beside the skeletal pelagic component is the mineralogy of the pellets. Similar to the bulk sediment, the mineralogy in tropical pellets is dominated by aragonite versus calcite at IODP site U1460. Tropical peloids are cemented by aragonite, while pellets at IODP site U1460 show HMC and dolomite cements.

A clear difference is also visible in the isotopic signature of the faecal pellets. The different carbon isotope values can partly be explained by mineralogical differences. Rubinson and Clayton (1969) have shown that the enrichment of δ13C in aragonite is about 1.8‰ higher compared to LMC. The aragonite content is about 80% higher at the tropical sites (Gischler, 2006; Gischler & Lomando, 2005) compared to IODP site U1460. The different aragonite content therefore could account for ~1.44‰ of the lower δ13C values seen in the pellets at IODP site U1460. However, the observed difference is on the order of ~3‰, indicating that aragonite concentration alone cannot explain the lower δ13C. Therefore, the higher δ13C values in tropical peloids might at least partially reflect the isotopic enrichment of the dissolved inorganic carbon pool due to photosynthesis observed in environments with a low water exchange rate (Swart & Eberli, 2005; Swart, Reijmer, & Otto, 2009). Carbonate cement precipitated under the influence of BSR, assumed to occur with the studied pellets, can show δ13C depletion (Machel, 2001). However, this effect is typically restricted by the buffering effect of the carbonate matrix and therefore is less important in this case. The most likely reason for the difference in δ13C is the higher content of skeletal material in pellets at IODP site U1460 that typically form in isotopic disequilibrium and generally have lower δ13C values (Gischler et al., 2009; Swart et al., 2009). The discrepancy in the δ18O signature between the pellets from the tropical sites and IODP site U1460 can be explained by difference in salinity and temperature. Temperatures and salinities are more elevated on the carbonate ramp of Kuwait (39‰–43‰, 13–32°C; Gischler & Lomando, 2005) and the carbonate platforms of the Belize-Yucatan system (35‰–42‰, 27–28°C; Gischler, Hauser, Heinrich, & Scheitel, 2003) compared to the environmental conditions on the outer ramp and slope of the Southwest Shelf of Australia (35.6‰–35.8‰, 17–18°C; James et al., 1999). Assuming a δ18O–temperature relationship of 0.21 ‰/°C (Bemis, Spero, Bijma, & Lea, 1998), the different δ18O values between IODP site U1460 and the Belize-Yucatan Platforms (~2.2 ‰) would indicate ~11°C lower temperatures at IODP site U1460. The match between these calculated and observed temperature differences indicates that much of the bioclastic sand and mud incorporated in the pellets was formed and cemented in relatively cool-bottom waters. This is also consistent with the similarity to the isotopic signature of HMC cements formed on the sea floor of the cool-water Lacepede Shelf (Figure 7). Higher δ18O water values in the arid, hypersaline setting of the Kuwait ramp likely explain the lower δ18O difference between pellets from that site and IODP site U1460. Evaporation leads to the preferential removal of 16O from surface waters and hence to an increase in δ18O. This is often termed the ‘salinity effect’, since the salinity is affected by the same processes as the isotopic fractionation in sea water. The observed offset in carbon and oxygen isotope values between IODP site U1460 and the modern examples is stable over the investigated interval (upper ~40 m; ~500 kyr; Figure 7) and therefore does not appear to be affected by early burial diagenesis. During deeper burial, the δ18O signature might be modified by further cementation and recrystallization potentially overprinting the primary isotopic differences (Reuning, Reijmer, & Betzler, 2002), but the δ13C signature is expected to be less prone to alteration (Reuning, Reijmer, & Mattioli, 2006; Reuning et al., 2002). Besides the planktic component, the isotopic signature therefore seems to be a criterion that can be used to differentiate shallow water tropical peloids from those deposited in deeper, and therefore cooler, waters.

6.3 Lithification of pellets

Cementation of pellets in shallow, tropical settings is controlled by elevated alkalinity and temperature, such as found on the Bahama Banks (Bathurst, 1975; Harris, 1979; Lees, 1975; Rankey & Reeder, 2010), north-western Arabian-Persian Gulf (Gischler & Lomando, 2005), Shark Bay (Logan, 1970; Read, 1974) and on the north-western shelf of Australia (Gallagher et al., 2018; James et al., 2004) and Bora Bora (Gischler, 2011). In the studied interval, the observation of frambooidal pyrite and aragonite dissolution at the shallowest occurrence of pellets (1.15 m CSF-A) indicates BSR in the near sea floor environment (Figures 6 & 7). During BSR, the sulphate (SO4 2−) is microbially reduced to hydrogen sulphide (H2S). The produced H2S is subsequently precipitated as iron sulphide (pyrite) (Biehl, Reuning, Schoenherr, Lüders, & Kulka, 2016; Goldhaber & Kaplan, 1974; Jorgensen, 1977). This reaction provides the acid required for aragonite dissolution (Wright, Cherns, Azerêdo, & Cabral, 2018), resulting in an increase in alkalinity that promotes the precipitation of calcite and dolomite cements (Canfield & Raiswell, 1991; Ku, Walter, Coleman, Blake, & Martini, 1999; Reuning et al., 2002, 2006). Additionally, the downwelling of saline waters derived from the restricted Shark Bay onto the outer ramp (James et al., 1999) likely provided higher alkaline waters that could have contributed to
early cementation of pellets in this area. A fossil example for the importance of very early cementation in the consolidation of faecal pellets is given by Friis (1995). He demonstrated that faecal pellets were only preserved in Tertiary siliciclastic sediments if they were stabilized by very early pyrite and calcite cements. Similarly, it is suggested here that very early cementation by calcite, dolomite and pyrite related to aragonite dissolution and BSR were essential for the preservation of the faecal pellets after burial.

7 | CONCLUSIONS

Faecal pellets can be as abundant in deeper water, subtropical environments as in shallow, tropical settings. On the Carnarvon ramp, faecal pellets are formed and preserved at the transition between the outer ramp/shelf and slope at a depth >200–300 m. They occur more abundantly during interglacials, when IODP site U1460 was situated in deeper waters below the swell wave base. Faecal pellets from deeper, cool-water carbonate settings are characterized by lower δ13C values and higher δ18O, and by higher planktic components compared to shallow water tropical peloids. Furthermore, the fact that they are composed of planktic and benthic components differentiates them from pellets produced by pelagic organisms. The near sea floor cementation by calcite, dolomite and pyrite related to BSR-driven aragonite dissolution is essential for the hardening of faecal pellets and their preservation in the fossil record.

ACKNOWLEDGEMENTS

We thank our two reviewers Laura G. O’Connell and C. Pederson for their constructive and detailed comments on our manuscript. The Integrated Ocean Drilling Program (IODP) is gratefully acknowledged for providing core samples for the study. Philipp Binger is thanked for his assistance during sampling in the laboratory. We gratefully acknowledge Dr. Schiebel for discussion. Prof. Dr. Thomas Brachert is thanked for providing the stable isotopes analyses. The authors also thank Uwe Wollenberg for his help with XRD measurements and SEM and EDX samples. SEPM is acknowledged for the permission to use parts of figures from James et al. (1999) for our Figure 1. This work was supported by a grant from the German Academic Scholarship Foundation (Studienstiftung des Deutschen Volkes) awarded to Hanaa Deik. Further funding was provided by the DFG (German Science Foundation, Project 320220579) to Lars Reuning.

CONFLICT OF INTEREST STATEMENT

The authors declare that there is no conflict of interest that could be perceived as prejudicing the impartiality of the research reported.

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REFERENCES

Alexandersson, E. T. (1979). Marine maceration of skeletal carbonates in the Skagerrak, North Sea. Sedimentology, 26, 845–852. https://doi.org/10.1111/j.1365-3091.1979.tb00977.x
Bandel, K. (1974). Faecal pellets of Amphineura and Prosobranchia (Mollusca) from the Caribbean coast of Columbia, South America. Senckenbergiana Maritima, 6, 1–31.
Bathurst, R. G. C. (1975). Carbonate sediments and their diagenesis. Developments in sedimentology, 12, (439 pp). Amsterdam, The Netherlands: Elsevier.
Bemis, B. E., Spero, H. J., Bijma, J., & Lea, D. W. (1998). Reevaluation of the oxygen isotopic composition of planktic foraminifera: Experimental results and revised paleotemperature equations. Paleoceanography, 13, 150–160. https://doi.org/10.1029/98PA00070
Betzler, C., Saxena, S., Swart, P. K., Isen, A., & James, N. P. (2005). Cool-water carbonate sedimentology and eustasy; Pleistocene upper slope environments, Great Australian Bight (Site 1127, ODP Leg 182). Sedimentary Geology, 175, 169–188. https://doi.org/10.1016/j.sedgeo.2004.11.008
Biehl, B. C., Reuning, L., Schoenherr, J., Lüders, V., & Kukla, P. A. (2016). Impacts of hydrothermal dolomitization and thermochemical sulfate reduction on secondary porosity creation in deeply buried carbonates: A case study from the Lower Saxonian basin, northwest Germany. American Association of Petroleum Geologists Bulletin, 100, 597–621. https://doi.org/10.1306/01141615055
Blom, W. M., & Alsop, D. B. (1988). Carbonate mud sedimentation on a temperate shelf. Bass Basin, southeastern Australia. Sedimentary Geology, 60, 269–280. https://doi.org/10.1016/0037-0738(88)90124-8
Brooke, B., Nichol, S., Hughes, M., McArthur, M., Anderson, T., Przeslawski, R., … Doherty, P. (2009). Carnarvon Shelf survey post-survey report. Geoscience Australia Record, 2, 90.
Canfield, D. E., & Raiswell, R. (1991). Carbonate precipitation and dissolution: Its relevance to fossil preservation. In P. A. Allison & D. E. G. Briggs (Eds.), Taphonomy: Releasing the data locked in the fossil record, (Vol. 9, pp. 411–453). New York, NY: Plenum Press.
Christensen, B. A., Renema, W., Henderiks, J., De Vleeschouwer, D., Groeneveld, J., Castañeda, I. S., … McHugh, C. M. (2017). Indonesian Throughflow drove Australian climate from humid Pliocene to arid Pleistocene. Geophysical Research Letters, 44, 6914–6925. https://doi.org/10.1002/2017GL072977
Collins, L. B., James, N. P., & Bone, Y. (2014). Carbonate shelf sediments of the western continental margin of Australia. In F. L. Chiocci & A. R. Chivas (Eds.), Continental shelves of the world: Their evolution during the Last Glacial-Eustatic Cycle (Vol. 41, pp. 255–272). London, UK: Memoirs, Geological Society. https://doi.org/10.1144/m41.19
Enos, P. (1974). Surface sediment facies map of the Florida Bahama Plateau. Geol. Soc. Amer. Map Series, MC-5(4th ed.). Boulder, CO: Geological Society of America.
Farrow, G. E., & Fyfe, J. A. (1988). Bioerosion and carbonate mud production on high-latitude shelves. Sedimentary Geology, 60, 281–297. https://doi.org/10.1016/0037-0738(88)90125-X
Friis, H. (1995). The role of fecal pellets in deposition of marine muddy sediments–examples for Danish Tertiary. Bulletin of the Geological Society of America Denmark, 42, 68–73.
Milliman, J. D. (1974). Carbonate sedimentation on Hogsty Reef, a Bahamian atoll. *Journal of Sedimentary Petrology, 37*, 658–676.

Milliman, J. D. (1969). Carbonate sedimentation on four southwestern Caribbean atolls and its relation to the “oolite problem”. *Transactions Gulf Coast Association of Geological Societies, Transactions*, 19, 195–206.

Milliman, J. D. (1974). *Marine carbonates* (p. 375). Berlin, Germany: Springer.

Nelson, C. S. (1988). An introductory perspective on non-tropical shelf carbonates. *Sedimentary Geology, 60*, 3–12. https://doi.org/10.1016/0037-0738(88)90108-X

Neumann, A. C., & Land, L. S. (1975). Lime mud deposition and calcareous algae in the Bight of Abaco, Bahamas: A budget. *Journal of Sedimentary Research, 45*, 763–786. https://doi.org/10.1306/212FeE3D-2B24-11D7-8648000102C1865D

Newell, N. D., Imbrie, J., Purdy, E. G., & Thurber, D. L. (1959). Organism communities and bottom facies, Great Bahama Bank. *Bulletin of the American Museum of Natural History, 117*, 177–228. http://hdl.handle.net/2246/1971

O’Connell, L. G., & James, N. P. (2015). Composition and genesis of temperate, shallow-marine carbonate muds: Spencer Gulf, South Australia. *Journal of Sedimentary Research, 85*, 1275–1291. https://doi.org/10.2110/jsr.2015.73

Petrick, B., Martínez-García, A., Auer, G., Reuning, L., Deik, H., Takayanagi, H.,… Haug, G.H.(2018). Glacial Indonesian Throughflow weakening across the Mid-Pleistocene Climatic Transition. Poster session presented at the AGU Fall Meeting, Washington.

Purdy, E. G. (1963). Recent calcium carbonate facies of the Great Bahama Bank, 1. Petrography and reaction groups. *The Journal of Geology, 71*, 334–355.

Rahimpour-Bonab, H., Bone, Y., Moussavi-Harami, R., & Turnbull, K. (1997). Geochemical comparisons of modern cool-water calcareous biota, Laceypede Shelf, south Australia, with their tropical counterparts. In N. P. James & J. A. D. Clarke (Eds.), *Cool-water carbonates: Shelf, Southern Australia*, (Vol. 56, pp. 77–91). Tulsa, OK: Society for Sedimentary Geology, Special Publication.

Ranken, E. C., & Reeder, S. L. (2010). Controls on platform-scale patterns of surface sediments, shallow Holocene platforms, Baham. *Sedimentology, 57*, 1545–1565. https://doi.org/10.1111/j.1365-3091.2010.01155.x

Read, J. F. (1974). Carbonate bank and wave-built platform sedimentation, Edel Province, Shark Bay, Western Australia. In B. W. Logan, J. F Read, G. M. Hagen, P. Hoffman, R. G. Brown, P. J. Woods, & C. D. Gebelein (Eds.), *Evolution and diagenesis of Quaternary carbonate sequences, Shark Bay, Western Australia* (Vol. 22, pp. 1–60). American Association of Petroleum Geologists, Memoir.