Sea-surface temperatures for the last 7200 years from the eastern Sunda Shelf, South China Sea: climatic inferences from planktonic foraminiferal Mg/Ca ratios.

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Abstract

To test whether low latitude shallow shelf deposits can provide high resolution paleoclimatic records, we utilized two cores from the Holocene sedimentary fill of incised valleys on the Sunda Shelf off Sarawak, Malaysia. We developed a new sea surface temperature (SST) record based on planktonic foraminiferal Mg/Ca for the last 7200 years. This record reveals several significant shifts between warmer and colder conditions. Temperatures averaged 27.5°C ca. 7200 cal. years BP, then climbed to 28.2°C from 6500-5500 cal. years BP. At 5500-4500 cal. years BP we identified the coldest period (26.8 °C) of the analyzed period. For the last 4500 years SST again averaged 27.5°C but the profile is rather variable. The last ca. 1000 years recorded the warmest SST averaging 28.5°C. We record, for the first time in this region, a cool interval, ca. 1000 years in duration, centered on 5000 cal years BP concomitant with a wet period recorded in Borneo. The record also reflects a warm interval from ca. 1000 to 500 cal years BP that may represent the Medieval Climate Anomaly.

Variations in the East Asian Monsoon (EAM) and solar activity are considered as potential drivers of SST trends. However, hydrology changes related to the El Niño-Southern Oscillation (ENSO) variability, shifts of the Western Pacific Warm Pool and migration of the Intertropical Convergence Zone are more likely to have impacted our SST temporal trend. Our findings indicate that climatic patterns in the region might be in phase with ENSO and out of phase with the EAM.

Keywords

Southeastern Asia; Sunda Shelf; East Asian Monsoon; paleoclimatology; Holocene; Foraminifera; Mg/Ca

1. Introduction

Geological reconstructions of climatic variations are essential for understanding current and future climate changes. They can extend the instrumental record back in time and provide information in geographic areas lacking comprehensive monitoring programs. In fact, instrumental data are geographically limited and provide accurate information only since ca. 1850 CE (e.g., IPCC, 2013). The development of proxy-based climatic reconstructions is most pressing considering the magnitude of global warming over the past ca. 150 years (e.g., Hartmann et al., 2013) and the potential climatic scenarios that indicate a likely temperature increase of over 1.5°C by 2081-2100 CE (Collins et al., 2013). Many Holocene climatic reconstructions are based on terrestrial records from extra-tropical or high-elevation sites (e.g., Mayewski et al., 2004; Wanner et al., 2008); global average surface temperature changes, however, closely follow those of the tropics (Oppo et al., 2009), which contain sedimentary sequences that may yield excellent climatic records. Holocene records derived from marine sediments tend to be from deep ocean waters (e.g., Ellison et al., 2006) and coral reefs (e.g., Beck et al., 1997). The scarcity of these records in certain locations leaves a limited understanding of Holocene climatic variations at a regional scale. The region affected by the East Asian Monsoon (EAM) has received considerable attention but the data have largely been derived from terrestrial and deep northern South China Sea (NSCS) records (Figure 1). However, Holocene climatic variations may also be recorded in coastal and shallow shelf environments, if sedimentation rates are large enough to provide the required temporal resolution and if sediment reworking is minimal (e.g., Kong et al., 2014). We address this unexploited source of paleoclimatic data in this paper.

While the Asian and Australasian monsoons impact half of the world's population, their patterns are difficult to predict and model, adding uncertainty to future climatic variation (Steinke et al., 2014). It is, therefore, imperative to provide complementary paleoclimatic data that expand the temporal reach of instrumental data and that are strategically distributed geographically. The monsoon system is a critical part of global atmospheric circulation that controls the redistribution of heat. In the SCS, the EAM influences wind patterns, precipitation, sea circulation, continental runoff, and nutrient transport (Huang et al., 1997; Cheng et al., 2005). The EAM is characterized by seasonal reversals, with summer (May–August) and winter (October–March) monsoons (Yoshino, 1971; An et al., 1991). The East Asian Winter Monsoon (EAWM) drives cold, dry winds from the northeast and is linked to the Indonesian Australian Summer Monsoon (IASM). In contrast, the East Asian Summer Monsoon

(EASM) drives southwesterly winds, bringing moist, tropical air to the region (Ding, 1994). These seasonal changes lead to strong seasonality in sea surface temperature (SST), salinity (SSS), and circulation patterns, among other parameters (Steinke et al., 2010).

We hypothesize that shallow continental shelf settings of the SCS can provide high resolution reconstructions of climatic proxies. To test this, we provide the first high-resolution reconstruction of Mg/Ca-derived sea surface temperatures for the Sunda Shelf region of the southern SCS (SSCS) for the last 7200 years (Figure 1). We utilize the planktonic foraminifera *Globigerinoides ruber* (d'Orbigny) and *Globigerinoides sacculifer* (Brady) from two cores in the Holocene sedimentary fill of two roughly southeast-northwest trending incised valleys off the coast of Bintulu, Sarawak, Malaysia (Figure 1).

The Sunda Shelf, occupying the southwestern, semi-enclosed SSCS, has limited connectivity with the Indian Ocean (through the Malacca Strait), the Java Sea (through the Gaspar and Karimata Straits), and the Pacific Ocean through the Sulu Sea, north of Borneo. The shelf is bounded by the Malay Peninsula, the southern Indo-China peninsula, Sumatra, Java and Borneo (Figure 1). It is extremely low-gradient (1:9,000), approximately 800 km in width, and averages ~70 m in depth (Hanebuth and Stattegger, 2003). Average salinity on the Sunda Shelf is ca. 33 (Zweng et al., 2013), although salinity values near the outflows of the major rivers draining Borneo can be much lower at the end of the rainy season (Wyrski, 1961). Sea surface temperatures range from approximately 26.5–28 °C during the EAWM to 28–29.5°C during the EASM (Locarnini et al., 2013). Holocene sediments on the shelf reflect sediment starvation and are generally thin (~1 m; Hanebuth et al., 2002, 2011), hence our targeting of incised valley-fill where there is greater accommodation space than on the adjacent shelf.

2. Materials and methods

Piston cores D42 (3.726500° N, 112.469967° E, 3.88 m long, 52 m water depth) and D45 (3.810183° N, 112.727200° E, 2.68 m long, 63 m water depth) were collected by the Department of Minerals and Geoscience Malaysia in the marine sedimentary fill of two incised valleys on the Sunda Shelf approximately 80 km northwest of Bintulu, Sarawak, Malaysia in July of 2008 (Figure 1). In January 2013, these cores were subsampled at contiguous, 2-cm intervals.

2.1 Mg/Ca analysis and temperature calculations

Between 200 and 450 micrograms of the shallow-dwelling (<50 m; Fairbanks et al., 1980, 1982; Mulitza et al., 1997; Anand et al., 2003; Farmer et al., 2007) planktonic foraminifera *G. ruber* and *G. sacculifer* were used for Mg/Ca analysis. The tests of foraminifera were picked from the >125 to 250 μm fraction of each sample after wet sieving. The mud to sand ratio was determined following sieving. Species identifications were verified via comparison with specimens housed at the Smithsonian Institution, Washington, D.C. The samples were cleaned following standard procedures (Boyle, 1981), excluding the reductive step, to remove clay and organic matter. Mg and Ca values were measured on a Jobin Yvon Ultima inductively coupled plasma atomic emission spectrophotometer (ICP-AES) at the University of South Carolina. To address the need for the reductive treatment, we used 10 samples from a nearby core (data not shown). In this test we split each of the 10 samples into two replicates and in one we applied the reductive treatment while the other was analyzed without the reductive step. The results indicated that our samples were adequately cleaned without the reductive step as indicated by very low Mn/Ca ratios in both sets and, therefore, we concluded that this step was unnecessary for our samples.

Following Barker et al. (2003) recommendations, the modern database to calibrate our samples should follow a similar cleaning procedure to minimize errors. Therefore we used the published dataset from Regenberg et al. (2009) that also eliminated the reductive step. This approach was necessary since the reductive method affects the Mg/Ca ratio by lowering it by ca. 15% (and therefore affects temperature calculations) as reported by Barker et al. (2003). Regenberg et al. (2009) provided a warm water, multi-species Mg/Ca-temperature relationship. However, to better represent our core material, we developed a new exponential relationship following the Hamilton and Shennan (2005) suggestion of developing training sets that best represent the target materials. A subset of modern samples from Regenberg et al. (2009; n=142), which included only *G. ruber* and *G. sacculifer*, was selected. Subsequently, sample Mg/Ca-temperature values were weighted to minimize bias due to uneven sampling along the temperature range analyzed (Kotze et al., 2012). The final Mg/Ca-temperature relationship based on *G. ruber* and *G. sacculifer* is indicated below, with a conservative error calculation of ± 0.64 $^{\circ}\text{C}$:

$$\text{Mg/Ca} = 0.1197e^{0.1391T}$$

Samples with Mg/Ca ratios larger than $6.0 \text{ mmol mol}^{-1}$ did not have modern analogues and samples with ratios larger than 8 mmol mol^{-1} were deemed unreliable (Figure 2).

2.2 Radiocarbon Age Estimates

Accelerator mass spectrometer (AMS) radiocarbon dating of the benthic foraminifer *Cavarotalia annectens* (Parker and Jones) provided a geochronological framework. Eight samples from D42 and 21 samples from D45 were submitted to NOSAMS (the National Ocean Sciences Accelerator Mass Spectrometry Facility) or Beta Analytic for AMS ^{14}C age analysis. The R/Bchron package (Haslett and Parnell, 2008; R Development Core Team, 2011), a Bayesian age-depth model, was used to calibrate radiocarbon age estimates (Marine09 calibration curve; Reimer et al., 2009) and generate age-depth models. Calibrated ages were compared with age estimates produced by Calib 7.0 using the Marine13 calibration curve (Reimer et al., 2013; Table 1) and corrected for the regional reservoir effect ($\Delta t \pm R$ of -38 ± 30 years). No significant differences were identified.

3. Results

3.1 Geochronology

Twenty-nine discrete AMS radiocarbon age estimates of the benthic foraminifer *Cavarotalia annectens* provide chronologic control for cores D42 and D45 and are provided in Table 1; the chronology for the complete cores was developed with R/Bchron and used in Figures 3 and 4. Three samples from D45 (3890, 2820, and 3050 ^{14}C years BP at 107 cm, 91 cm, and 79 cm, respectively) were excluded from the age-depth model because these ages were older than those preceding them stratigraphically and were reported as outliers by the age-depth model. Thus, they were considered to be the result of reworking or contamination. Similarly, samples below 200 cm in D42 were not considered for this reconstruction since the three lowermost samples yielded similar ages, although they were from different depths. From these chronologies, an average sedimentation rate of ca. 0.068 cm yr^{-1} in D45 and 0.156 cm yr^{-1} in the top 2 m of D42 can be calculated, providing a resolution of ca. 30 and 13 years, respectively, for each 2 cm sample. However, these rates were variable and some sections of these cores yielded a resolution of ca. 100 years per sample.

3.2 Sedimentology

Cores D42 and D45 comprise $97 \pm 2.6\%$ and $98 \pm 0.6\%$ (mean $\pm$ SD), respectively, gray mud. A thin (18 cm) unit around 3 m depth in D45 has lower mud content ($87 \pm 2.4\%$). Sand-sized particles are composed of whole and fragmented biogenic material including foraminifera, gastropods, bivalves, sponge spicules. The bottom 18 cm of D45 are barren of shell material and are composed of mud with some fine plant fragments.

3.3 Mg/Ca ratios and Sea Surface Temperature Records

Mg/Ca ratios from *G. ruber* and *G. sacculifer* range from 4.1 to 14.1, with an average value of 5.8 mmol mol⁻¹ (Figure 3). After removing the samples with a ratio larger than 8 mmol mol⁻¹, the average is 5.6. Ratios of modern analogues range from 3.0 to 6.0 mmol mol⁻¹. For the period of time during which D42 and D45 overlap, they exhibit similar mean values (5.98±0.88 and 5.64±0.95 mmol mol⁻¹, respectively) and were combined into a composite record (Figure 3C) after exclusion of eight outliers (>8 mmol mol⁻¹).

Reconstructed temperatures in D42 range from 25.7 to 31.1°C with an average value of 27.7°C, comparable to D45. However, D45 reconstructions reached up to 34.3°C (Figure 3B). When the eight outliers are removed, both cores exhibit very similar averages of ca. 27.5°C (ranging from ca. 25.4 to 30.2°C). A three-point moving average of the combined data set (D42 and D45) displays an average SST of 27.5°C (ranging from 25.6 to 29.4°C) for the last 7200 years (Figure 3C), and reveals several significant shifts between warmer and colder conditions. Temperatures were initially cold ca. 7200 cal years BP, then climbed to 28.2±0.5°C (median±SD) from 6500–5500 cal. years BP. This was followed by a colder period lasting ca. 1000 years, centered at 5000 cal. years BP (26.8±0.4°C). From 4500 to 1100 cal. years BP, the reconstructed temperatures are variable (27.5±0.5°C) but warm to 28.5±0.5°C for an interval of ca. 1000 years (boxplot on Figure 3D), at the top of the core.

4. Discussion

4.1 Suitability of the reconstruction, modern analogues, calibration equation, errors and validation

This work is based on the assumption that planktonic foraminifera are autochthonous (i.e., live in the inner shelf) and, therefore, that they represent the region's climatic patterns. Live planktonic foraminifera in shallow inner shelf waters have been recorded since at least the 1960s; Lidz (1966) recorded live planktonics in plankton tows (from 15 m deep waters) off southern California over waters from nearshore to ~100 m depth. More recent papers (e.g., Mohan et al. 2013) record live planktonics from tows in bays off the Andaman Islands. On the Sunda Shelf, patterns of percent planktonics in foraminiferal assemblages suggest that planktonic foraminifera do not live in waters <30 m deep but do live at depths >30 meters and their relative abundance in foraminiferal assemblages increases towards deeper environments (Biswas, 1976; Tu and Zheng, 1991; Szarek et al., 2006; Culver, Leorri et al.,

unpublished data on the west side of the South China Sea off peninsular Malaysia). In addition, this region experiences intense rainfall, resulting in a significant amount of river discharge onto the shallow shelf. This should be reflected in seaward sediment transport, a process concentrated in incised valleys such as the ones studied here. This hydrodynamic control should minimize potential import of allochthonous planktonic foraminifera into the region from open waters. Further, the study area is 150 to 200 km from the shelf break and, thus, upwelling is unlikely to be a factor.

In order to obtain enough material we combined two species to perform the analyses and we discuss below the potential source of error derived from using a multi-proxy approach. While *G. ruber* inhabits the top ~30 m of the water column (Fairbanks et al., 1980, 1982; Mulitza et al., 1997) whereas *G. sacculifer* calcifies at ~25–40 m depth (Farmer et al., 2007); thus the two species might reflect water conditions that are slightly different; although Anand et al. (2003) suggested a calcification depth for both species of 0–50 m. However, if as indicated above these species live in the water column in the region of the study area and are not transported from offshore, the well mixed waters (mixed layer depth between 35 and 55 m; Qu et al., 2007; Steinke et al., 2010) over the coring locations (50 to 60 m water depth) would result in minimal differences, if any, in data acquired from these two taxa. Additionally, similar multi-species calibrations have been performed previously (e.g., Elderfield and Ganssen, 2000; Anand et al., 2003; Regenberg et al., 2009). Regenberg et al. (2009) suggested that uncertainties in Mg/Ca calculated temperature amount to ± 0.24 °C at 19 °C and ± 0.29 °C at 28 °C, with an average error of ± 0.26 °C for a calibration curve based on four species from a temperature range of 19 to 28 °C. We used the same calibration dataset but limited our analysis to two species and a smaller temperature range, which would reduce rather than increase any reconstruction errors. The average maximum bias introduced by using two species is 0.77°C (based on the Regenberg et al., 2009 data set), which could partially explain the serrated pattern. However, only in 14 instances in our record samples alternate from both species to an adjacent sample that is monospecific and the peaks in our data are not associated with these changes, except on one occasion. It seems more plausible that differences in the *G. ruber*/*G. sacculifer* ratio could create some noise, but it does not seem to be responsible for the large variability. The potential error derived from the use of two species is captured in our training data set and included in the calibration curve error, which in the current study is ± 0.64 °C.

Foraminiferal test size impacts Mg/Ca ratios and generally has lower ratios for smaller tests (Elderfield et al., 2002). Test size in our cored samples is smaller (125 to 250 microns) than that analyzed by Regenberg et al.

(2009). Thus, our results might be biased towards lower Mg/Ca values, but the samples within both cores should be comparable (all of them are equally impacted by the size effect). Inter-sample variability due to test size (using the data provided by Elderfield et al., 2002) is less than 0.21 mmol mol⁻¹ for *G. sacculifer* and 0.28 mmol mol⁻¹ for *G. ruber*. It could be as large as 0.5 mmol mol⁻¹ for the latter taxon based on one sample that here was considered an outlier, which could invoke a maximum error of 0.2 to 0.3 °C at 26 °C. The relatively small test size of these two species could be a reflection of their living outside their environmental optima and that the shallow shelf is not a favorable habitat for these organisms (e.g., Schmidt et al., 2004). At the same time, it reinforces the argument that they are autochthonous.

Foraminiferal Mg/Ca-derived paleothermometry is widely used for paleoceanographic reconstructions and is based on the exponential relationship between foraminiferal Mg/Ca and sea water temperature (Nürnberg et al., 1996). However, Mg/Ca-SST calibrations differ significantly among studies (see Sadekov et al., 2008 and references therein). In addition to species-specific calibrations, several studies have focused on multi-species calibrations (see Regenberg et al., 2009 and references therein). While there are limitations to Mg/Ca-derived paleothermometry calculations, including sample size, analytical errors (ca. 0.1 mmol mol⁻¹), “vital effects”, pH, salinity, seasonality, habitat, etc. (Zeebe, 1999; Russell et al., 2004; Sadekov et al., 2008), reconstruction errors are only ca. 1.1°C (Anand et al., 2003), and less than 1.0°C if the sample size is adequate (Sadekov et al., 2008). Errors for individual samples are as low as 0.8°C (Regenberg et al., 2009) for *G. ruber* and *G. sacculifer*. Combining data from 142 samples of *G. ruber* and *G. sacculifer* included in the Regenberg et al. (2009) dissolution-corrected dataset, we adapted the multi-species exponential relationship to better capture samples with higher Mg/Ca ratios and, thus, reflect data from the cores of this study. Rosenthal et al. (2003) suggested that open ocean calibrations might not be appropriate for studies involving shelf foraminifera because high-Mg calcite is preferentially dissolved, yielding lower than expected SST estimates. The Regenberg et al. (2009) correction for the dissolution effect minimizes this problem.

Figure 6 in Regenberg et al. (2009) presents multispecies (eight species) Mg/Ca data versus calibrated calcification temperatures. From this figure it can be inferred that there is a significant change in the relationship between Mg/Ca and $\delta^{18}\text{O}$ -calcification temperature above 4.5 mmol mol⁻¹. With our newly developed calibration equation, this change is better captured, producing a more accurate calibration curve for the temperature range studied here. The error resulting from the newly developed equation is $\pm 0.64^\circ\text{C}$ (1 sigma). The sample specific total

error calculation follows Shennan and Horton (2002; total error= $\sqrt{(e_1^2+e_2^2+...e_n^2)}$) and includes the sum of analytical error, sample size bias, and calibration curve error, equating to $\pm 0.69^\circ\text{C}$ (1 sigma).

It is important to note that the Regenberg et al. (2009) modern analogues did not exceed ca. 6 mmol mol^{-1} and, therefore, SST reconstructions from samples with higher ratios should be considered carefully. Despite this constraint, we will discuss the temperature profiles instead of the Mg/Ca ratios since the calibration does not change trends in the data; only the units and amplitude change (i.e., the core profiles based on Mg/Ca ratios or calibrated temperature are exactly the same; Figures 3 A and B). However, the temperature reconstructions allow us to compare our results with other temperature reconstructions and because almost 80% of our samples had modern analogues, meaningful comparisons can be achieved. Other published work has also expanded calibration curves beyond the range covered by the actual Mg/Ca values (e.g., Ferguson et al., 2008 revised data set from Elderfield and Ganssen, 2000). Regardless, samples from this study with ratios larger than 8 mmol mol^{-1} were discarded because simple extrapolation may not capture changes in the Mg/Ca relationship.

Specimens in this study are small (<250 micrometers), delicate, well-preserved, and do not exhibit visual evidence of overgrowth, although visual inspection might not be enough to identify high-Mg overgrowth. Thus, the elevated Mg/Ca values identified herein are unlikely to be a result of high-Mg overgrowths typically formed in CaCO_3 supersaturated pore water. High salinity is unlikely to be a contributing factor to elevated Mg/Ca values (Hertzberg and Schmidt, 2013) because extensive freshwater input to the Sunda Shelf results in generally lowered salinity relative to open ocean values (Linsley et al., 2010).

We conclude that the modern analogue dataset used here is suitable to reconstruct Holocene paleotemperature and provides a high-resolution SST reconstruction for the last 7200 years in the southern SCS (Sunda Shelf). However, from our analysis it can be inferred that the Mg/Ca-temperature relationship might change at certain thresholds (as indicated at ca. $4.5 \text{ mmol mol}^{-1}$). This needs to be explored further by including analysis of modern samples from regions with calcifying temperatures greater than 28°C .

4.2 Holocene reconstruction, regional comparison, and climatic inferences

Figures 3 A and B provide Mg/Ca and temperature profiles for cores D42 and D45. They present serrated profiles with large amplitude changes. Similarly, changes from sample to sample reach 1.63°C in contiguous samples even when using a single species (Steinke et al., 2008).

Average sedimentation rates of ca. 0.068 cm/yr in D45 (with significant inflections) and 0.156 cm/yr in the top 2 m of D42 are calculated from the age/depth model, with no evident interruptions that would explain the serrated profiles. Furthermore, the depositional environments where the cores were collected are dominated by mud (cores D42 and D45 comprise $97 \pm 2.6\%$ and $98 \pm 0.6\%$), reflecting low energy conditions; reworking is likely not as intense as in sand-dominated regions. Bioturbation, on the other hand, would have the opposite effect: the downward transport and vertical mixing of sediments would provide a more homogenous core with gradual changes. The Mg/Ca profiles in Figures 3A and B suggest little bioturbation and/or sediment mixing.

One plausible explanation for the serrated character of the record could be that the shallow waters of the study area are more sensitive to changes in precipitation and runoff, which would more significantly impact inner shelf water chemistry than more open waters further from land. Salinity changes in the upper 50 m of the water column on the Sunda Shelf respond to seasonality and fluctuate by 1.2 (at 10 m) and 0.5 (at 50 m) (Steinke et al., 2010). Closer to shore, changes are even more dramatic. At the mouth of the Rajang River delta (~400 km southwest of the study area), salinity changes from 10 during the wet season to 25 during the dry season, reflecting the change in river discharge (Staub et al., 2000). River discharge also affects pH values, lower for fresh water. Ferguson et al. (2008) suggested that Mg/Ca increases by 15-59% per salinity unit increase. Kisakürek et al. (2008) also suggested that salinity impacts Mg/Ca ratios (they incorporated salinity into their calibration curve) and that pH also impacts this ratio. Their experiment with variable pH yielded $3.1 \text{ mmol mol}^{-1}$ at a pH of 8.43 but $7.5 \text{ mmol mol}^{-1}$ at a pH of 7.91. In the region, sites located closer to the river discharging area would be most affected, while our study site should be less impacted by variability in river discharge changes, however, decadal to centennial trends could impact the overall results. For example, Stephens et al. (2008) reported rainfall patterns for the period 1996 to 2002 and observed a drought during El Niño-Southern Oscillation (ENSO) and a decrease in peaks in seasonal rainfall from mid-1998 to 2002.

In addition to these environmental explanations, natural variability could also contribute to the variability in our Mg/Ca record. For example, Ferguson et al. (2008) Figure 1 (modern dataset) shows *G. ruber* $>8 \text{ mmol mol}^{-1}$ at ca. 21°C and $>13 \text{ mmol mol}^{-1}$ at 24°C . These could be also related to seasonal temperature changes that are known to impact foraminiferal Mg/Ca ratios (up to 2 mmol/mol for *G. ruber*; Anand et al., 2003).

In order to minimize the impact of the variability in our interpretations, we combined the data from both cores and utilized a three-point moving average to produce a combined record of SST. This record still displays a

serrated pattern with some significant oscillations between 7200 and ca. 4000 years ago and an irregular but apparent warming trend for the last ca. 4000 years (Figure 3C). Temperatures climb from colder values (ca. 27°C) at 7200 cal years BP to ca. 28.2°C at 6000 cal years BP, then cool to less than 27°C at 5000 cal years BP. The sea then becomes warmer, reaching 28.5°C on average over the last ca. 1000 years (Figure 3C; 3D). This increase is more pronounced over the last 600 years during which SST remained at almost 29°C, on average. Note, however, that the last 1000 years are characterized by an oscillating pattern in SST and the inferences from this period are limited by the current chronology. Overall, the data suggest an increase of ca. 1.5°C from ca. 27°C at 7200 cal years BP to almost 29°C, on average, for the last 600 years. Nevertheless, this overall warming trend is characterized by large temperature oscillations (e.g., cooler at 5000 cal years BP and warmer after ca. 1000 cal years BP; Figure 3C; 3D) rather than a gradual temperature increase, and these oscillations might mask the longer term (mid- to late Holocene) cooling that other authors have suggested for the region (see below). Temperature changes are better reflected in Figure 3D where four periods are identified in the box-plot. The most interesting feature is the significant cooling event centered at ca. 5000 cal years BP that lasted 1000 years, the first time a cooling event of this age has been recorded in marine sediments from this region. The boxplot analysis (Figure 3D) also suggests that the period between 6500 and 5500 years ago was warmer than the period between 4500 and 1100 years ago.

Regionally, our three-point averaged composite curve (Figure 3C) overlaps significantly with the Mg/Ca-derived paleo-temperatures from the Bali Basin, Makassar Strait (Linsley et al., 2010; Figure 4C and F), except during the colder interval around 5000 cal years BP, that is only hinted at by the lower resolution data from Steinke et al. (2006). Most proxy records do not extend into the last 1000 years but the warmer period starting at 1100 cal years BP in our record coincides with reconstructions from SW Sulawesi, Makassar Strait (Linsley et al., 2010; Figure 4F) and the regional reconstruction by Linsley et al. (2010; Figure 4C).

However, correlation with lower resolution published studies from the SCS is difficult due to their varying results. Briefly, two records seem to agree with our reconstruction (Kienast et al., 2002; Steinke et al., 2008) and three others, some produced by the same authors, exhibit only moderate overlap (Kienast et al., 2002; Steinke et al., 2006, 2008; Oppo et al., 2009). Linsley et al. (2010) developed a comprehensive paleoclimatic regional reconstruction for the Holocene, including data from the SCS, although these data were not included in their final composite SST reconstruction. Their study suggests a gradual SST cooling from between 10,000 and 7000 years ago to immediately prior to the industrial period, interrupted by a warming during the Medieval Warm period and

cooling during the Little Ice Age. Comparison of our SST curve with the Linsley et al. (2010) regional reconstruction over the last 1000 years is limited by our chronology, although our record also seems to record the Medieval Warm Period and Little Ice Age (Figure 4C).

Considering the discussion above, data from the NSCS (Huang et al., 1997; Wang et al., 1999; Yu et al., 2005; Kong et al., 2014), and notwithstanding the previously discussed limitations for regional comparisons, regional data suggest a general cooling since at least ca. 7000 years ago until ca. 1000 years ago (Yu et al., 2005; Linsley et al., 2010; Kong et al., 2014). This pattern is supported by climatic studies from mainland south Asia that identified warm temperatures from 7000 to 6500 years ago (Chabangborn and Wohlfarth, 2014). None of those records identified the cooling event at ca. 5000 years ago present in our dataset, which suggests more complex dynamics off Sarawak, Malaysia. Below we discuss some potential drivers that could explain the oscillations observed in our record.

Borneo straddles the East Asian monsoon system (Dubois et al. 2014), which is recognized as the main control of circulation in the SCS (e.g., Fang et al., 1998; Hu et al., 2000). However, these authors divide the SCS into three sections: NSCS, SSCS, and the central region. This division seems to be in agreement with Peel et al. (2007) climatic classification in the region. In general, in the absence of major oceanic inflow, surface currents undergo seasonal reversals, responding to the EAWM/EASM. During the EAWM, cold winds and the southward flow of colder waters result in low SSTs in the NSCS and a strong north-south temperature gradient within the SCS accompanied by minimum salinity. On the other hand, during the EASM season, SSTs are uniformly high ($\sim 28^{\circ}\text{C}$ – 29°C) in the SCS together with higher salinity (Steinke et al., 2010). That said, while satellite SST data are available from the NSCS, this information is sparse from the SSCS (Hu et al., 2000).

In addition to the EAM, the NSCS is also controlled by water exchanges with the East China Sea and the Kuroshio Current, where additional factors must be considered, such as solar irradiation and the topography of the SCS (Hu et al., 2000). At the present, our understanding of the currents is still limited, and while the SSCS is better understood, the NSCS seems to have more complex dynamics (Hu et al., 2000). For instance, many studies are based on the surface layer (0–200 m) (Hu et al., 2000). In this situation, shallower regions might not be properly covered. For instance, exchange of waters between the SCS and the Sulu Sea was studied by Metzger and Hurlburt (1996) but the straits connecting the SCS and the Sulu Sea were smaller than the study's resolution. The authors discounted any significant exchange of waters, but this might limit our understanding of the region. Steinke et al.

(2001) reported similar annual average temperatures between our study area (within the SCS) and the Sulu Sea, where they were connected by the same isotherm line.

SST in the study area has often been linked to changes in the monsoon system. Records of monsoon variability are largely derived from records of speleothem oxygen isotopes, a proxy for monsoon rainfall on landmasses surrounding the SCS. Data from Dongge (southern China; Dykoski et al., 2005; Wang et al., 2005), Heshong (southwest China; Hu et al., 2008), Wanxiang (central China; Zhang et al., 2008), and Sanbao caves (central China; Dong et al., 2010) suggest EASM weakening during the Holocene related to decreasing summer insolation. More recent records by Selvaraj et al. (northeastern Taiwan; 2012) and Zhao et al. (a synthesis of marine and terrestrial data from the western Pacific and Indian Oceans; 2013) suggest that the EASM may have strengthened in the late Holocene. With regards to the IASM, data from Sulawasi (Indonesia; Tierney et al. 2012); Liang Luar Cave (Indonesia; Griffiths et al. 2010); and cave KNI-51 (Australia; Denniston et al. 2014) suggest a dynamic monsoon with relationship to the Pacific Walker Circulation and ENSO. Figure 4D compares the Dongge Cave $\delta^{18}\text{O}$ data (Wang et al., 2005), which represents the “regional picture of monsoon events over the wider IASM and EASM regions” (Denniston et al. 2013, 155), with our three-point averaged composite record. There seems to be no clear correlation. This should not be interpreted as a lack of influence on SST by the monsoon system but instead as a lack of a linear correlation. Increased rainfall (and the associated increase river discharge) could be reflected in lowered temperatures in coastal areas but be coupled with higher atmospheric temperature. The phase/anti-phase relationships between rainfall and temperature, both controlled by the EAM, could mask the SST/EAM correlation.

Variations in the strength of the Holocene Asian Monsoon have been correlated to solar activity (SA) (Wang et al., 2005; Steinhilber et al., 2012). In order to explore further the potential influence of the monsoon system in our SST record we analyzed its correlation with SA. Our three point averaged composite SST record, plotted with trends of total solar irradiance from Steinhilber et al. (2009; Figure 4E) visually suggests a negative correlation. Correlation analysis, however, indicates generally poor correlation over the 7200 year record with moderate correlation for particular periods of time. The general pattern could suggest a weak correlation where SA, and as a result the EAM, is a major driver, albeit overprinted by other factors. However, the poor correlation might be due to the difficulty in matching the temporal resolution of our SST records with solar irradiance data and/or the interference of local or regional signals (e.g., cloud cover, winds and currents, rainfall intensity, and river discharge).

Indeed, similar lack of correlation was found between solar activity and monsoon intensity ($\delta^{18}\text{O}$ from Dongge cave), between 6000 and 5000 years ago and over the last 1500 years (Steinhilber et al., 2012). Note that the discrepancy between climatic proxies and SA at 6000 to 5000 years BP was also identified in Alaska (Hu et al., 2003). The difficulty of correlating SA with climate data does not necessarily imply that SA should be discarded as a driver of regional SST. The Steinhilber et al. data (2012; see their figure 3B) indicate higher cosmic ray intensity ca. 7000 years ago, a generally decreasing trend until ca. 1500 year ago, followed by an increase in intensity. Between ca. 5500 and 4800 years ago, and during the last ca. 1500 years, oscillations in cosmic ray intensity became larger and more irregular. Approximately 5500 years ago, total solar irradiance reached minimum values for the past 9400 years. Steinke et al. (2010) suggested that the EAWM was weakest from 6500 to 2500 years ago, roughly in agreement with the cosmic ray intensity reconstruction. The SA trends (Steinhilber et al., 2012) are in general agreement with the regional cooling of SST (Linsley et al., 2010) and the SA oscillations are roughly coincident with the major excursions in our SST data.

The bottom of core D45 represents a relatively warm period around 6000 years ago (excluding the oldest four samples that could reflect sample variability; these samples occur immediately above a probable non-marine unit and may represent nearshore conditions) that, as an alternative to the EAM and SA as major drivers, could be explained by a westward shift or an expansion of the western Pacific warm pool (WPWP) during the Holocene climatic optimum (HCO) 10,000 to 7000 years ago (Linsley et al., 2010). Our reconstruction only captures the last part of the HCO. The WPWP migrated eastward during the mid- to late Holocene (Linsley et al., 2010), resulting in cooler temperatures in the SCS reflecting the smaller influence from the WPWP. In this case, our record would support the inferred WPWP expansion and/or shift during the early Holocene as proposed by Linsley et al. (2010). Note that the temperatures from our record are cooler for the period 5500 to 1100 than for 6000 years ago (Figure 3D). However, the oceanographic connectivity of our site with the WPWP might have been limited as indicated above.

The WPWP expansion and/or shift does not explain the significantly colder period in our record centered ca. 5000 years ago. However, anomalously high convective activity in the WPWP during the period between 3500 and 5500 cal years BP has been linked to a maximum in boreal fall insolation on the equator, an increase in tropical Pacific's SST gradient, and a relative wet period in Borneo (Partin et al., 2007; Chen et al., 2013). As we indicated above, our record might be more sensitive to changes in precipitation and runoff changes, and in fact, the coldest

period detected in our record coincides with a regional relatively wet period as reflected by $\delta^{18}\text{O}$ values in the Borneo stalagmite record (Partin et al., 2007; Figure 4B).

ENSO, which is considered the primary driver of inter-annual climate variability in the tropics and subtropics (Chen et al., 2016), also exerts the dominant control on north Borneo precipitation (Partin et al., 2007). At a monthly time scale, rainfall measured from 01 February 1996 to 22 April 2002 in north-central Sarawak shows the 1997–1998 El Niño drought (Stephens et al., 2008). At a millennial time scale, using data from Borneo, Chen et al. (2016) have recently connected the reduced interannual variability in ENSO with the anomalously high convective activity in the WPWP and the wet period in Borneo (negative excursion in stalagmite $\delta^{18}\text{O}$, Partin et al., 2007), although the chronology of these events (and other records from the east equatorial Pacific) is roughly similar (4000 to 6000 years ago) due to the complexities of the ENSO events and the sensitivity of the records. Chen et al. (2016) suggested that reduced ENSO variability might correlate with the boreal fall insolation and be out of phase with EASM variability. Stalagmite $\delta^{18}\text{O}$ from Borneo suggests that the cold period detected in our record at ca. 5000 years ago reflects the hydrology of the region and might be connected to the southward migration of the Intertropical Convergence Zone (ITCZ) during the mid-Holocene as a response to precessional forcing, crossing the equatorial Pacific and delivering more rainfall to the study area ca. 5000 years ago. The increased rainfall is recorded by shifting $\delta^{18}\text{O}$ values in the Borneo stalagmite record (Partin et al., 2007; Figure 4B) and further supported by pollen data (Chabangborn and Wohlfarth, 2014). This increase rainfall is further linked to a generally weaker variability of the ENSO during the 4000 to 6000 year BP period (Chen et al., 2016) and therefore the cold period identified in our core centered on 5000 cal years BP is of global significance.

The interval from ca. 1000 to 500 cal yr BP exhibits the highest temperatures ($28.2\pm0.5^\circ\text{C}$; maximum 29.3°C) in our record and is followed by a drop in temperature to 28.0°C ca. 300 cal yr BP (Figure 3). The warm interval could reflect the Medieval Climate Anomaly (MCA) and the cooler interval, the Little Ice Age (LIA). The MCA has been documented in several paleoclimate studies in the region (e.g., Newton et al., 2006, 2011; Rosenthal et al., 2013; Kong et al., 2014), and is also recognizable in the Linsley et al. (2010) SST reconstruction. However, to conclusively identify the MCA and the LIA on the Sunda Shelf, higher resolution sampling is required.

5. Conclusions

Mg/Ca data from two cores recovered from incised valleys off Sarawak, Malaysia, combined with 29 AMS C-14 age estimates and analyzed using a new Mg/Ca-temperature relationship derived from published data, provide a higher resolution SST record than is possible over the rest of the generally sediment starved Sunda Shelf. These findings alone are significant, as they demonstrate new, unexploited sediment sources for producing high resolution Holocene paleoclimatic records in the SSCS. When compared to other regional records, the magnitude of the SST oscillations and values are in moderate to good agreement with previously reported SSTs. Regionally, the general trend of SST is characterized by cooling from ca. 7000 to ca. 1000 years ago. In our record, the interval from ca. 1000 to 500 cal yr BP exhibits the highest temperatures followed by a temperature decrease by ca. 300 cal yr BP, in agreement with published records. However, our record also contrasts with several previous studies, but this may be due to Holocene temperature oscillations not recorded in lower resolution SST studies. Indeed, the newly developed data herein suggest a 1000-year cool period centered at ca. 5000 years ago, coeval with an increase in rainfall in Borneo. We explored the relationship between the variations in SST with SA or variations of the EAM but no clear correlation was evident. However, migration of the ITCZ connected to shifts of the WPWP, and weaker variability of the ENSO could explain this cool period. Our findings highlight the complexity of controls on Holocene climate on the Sunda Shelf and the need for additional SST records from shallow shelf environments.

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Figure 1. **Left:** A map of the Sunda Shelf and surrounding region in south-east Asia. The hashed rectangle indicates the area shown at right.

Right: Bathymetric map of the study area off Bintulu, Sarawak. Contours (m) were generated from data from chart MAL 751, published December 31, 1993 by the Royal Malaysian Navy. Black squares indicate core locations.

Figure 2. **Top:** *G. ruber* and *G. sacculifer* Mg/Ca vs. temperature calibrations based on data from Regenberg et al. (2009). The calibration curve is based on a weighted distribution of the samples based on temperature ranges (given by the analytical limitation, 0.1 mmol/mol) to minimize bias due to uneven sampling along the temperature range analyzed.

Bottom: Table indicating the ranges used to develop the calibration curve above.

Figure 3. **Panel A** represents Mg/Ca ratios for each core. **Panel B** represents the calculation of SST ($^{\circ}\text{C}$) based on the Mg/Ca ratios from Panel A using the equation developed herein. **Panel C:** Composite (D42 and D45) reconstruction of SST ($^{\circ}\text{C}$) with a 3-point average (to maximize the signal-to-noise ratio) over the last 7200 years. **Panel D:** box-plot of temperature ranges ($^{\circ}\text{C}$) for discrete time intervals (ka cal years BP) from the SST reconstruction in panel E.

Figure 4. **Panel A:** Composite (D42 and D45) reconstruction of SST ($^{\circ}\text{C}$) with a 3-point average (to maximize the signal-to-noise ratio) over the last 7200 years. **Panel B:** $\delta^{18}\text{O}$ from Borneo (Partin et al., 2007). **Panel C:** Regional temperature anomaly ($^{\circ}\text{C}$) from Linsley et al. (2010). **Panel D:** Comparison between Panel A (SST, this study) and the Dongge Cave $\delta^{18}\text{O}$ record (Dykoski et al., 2005). **Panel E:** Comparison between Panel A (SST, this study) and total solar irradiance (TSI) (Steinhilber et al., 2009). Dashed boxes indicate potential correlation between records. **Panel F:** Comparison between SST ($^{\circ}\text{C}$) in panel E with the records from Core 13 GGC and Core 70 GGC (Linsley et al., 2010).