3500 yr record of centennial-scale climate variability from the Western Pacific Warm Pool

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ABSTRACT

We use geochemical data from a sediment core in the shallow-silled and intermittently dysoxic Kau Bay in Halmahera (Indonesia, lat 1°N, long 127.5°E) to reconstruct century-scale climate variability within the Western Pacific Warm Pool over the past ~3500 yr. Downcore variations in bulk sedimentary $\delta^{15}N$ appear to reflect century-scale variability in basin ventilation, attributed to changes in oceanographic conditions related to century-scale fluctuations in El Niño Southern Oscillation (ENSO). We infer an increase in century-scale El Niño activity beginning ca. 1700 yr B.P. with peaks in El Niño activity ca. 1500 yr B.P., 1150 yr B.P., and ca. 700 yr B.P. The Kau Bay results suggest that there was diminished ENSO amplitude or frequency, or a departure from El Niño–like conditions during the Medieval Warm Period, and distinctive, but steadily decreasing, El Niño activity during and after the Little Ice Age.

Keywords: climate, El Niño Southern Oscillation, Western Pacific Warm Pool, Indonesia.

INTRODUCTION

Kau Bay is a small (30 × 60 km), intermittently anoxic, ~470-m-deep basin that is semi-enclosed by the island of Halmahera (Indonesia, lat 1°N, long 127.5°E) and connected to the equatorial Pacific Ocean and Western Pacific Warm Pool (WPWP) by an ~30-m-deep, 15–20-km-wide sill (Fig. 1). Because water exchange is limited to the upper 30 m, Kau Bay's deep-water temperature and salinity are nearly homogeneous below the mixed layer and reflect the surface water hydrography outside the bay (Van Aken and Verbeek, 1988; Van der Weijden et al., 1989; Van Riel, 1943) (Figs. 2A–2D). The deep basin's dissolved oxygen concentrations vary and indicate intermittent ventilation (Van Aken and Verbeek, 1988; Middelburg, 1990). Middelburg (1990) estimated that the oxygen minimum zone observed in 1985 may have developed in ~120 days and that the 150-m-thick anoxic layer observed in 1930 developed in fewer than 3 yr (Figs. 2E and 2F).

Temperature and [O_2] (Fig. 2) of the upper ~20 m surface layer within Kau Bay reflect open ocean surface water values above the sill depth (Fig. 2). Freshwater input from the surrounding land reduces surface salinity within the bay relative to outside the bay, and stratifies the water column (Fig. 2). Ventilation of the entire water column within the basin may, however, occur when the wind- and current-driven flux of saltier and/or denser water entering the basin overcomes the salinity gradient, leading to deep mixing. Van Aken and Verbeek (1988) proposed that flushing of Kau Bay is possible annually during September–November, when the New Guinea Coastal Current (NGCC) introduces slightly higher salinity water to the vicinity of Kau Bay (Arruda and Nof, 2003; Masumoto et al., 2001; Wyrtki, 1961). The southward-flowing Mindanao Current (MC) and the seasonally northwest- and southeast-flowing NGCC collide near Halmahera to develop the cyclonic Mindanao Eddy (ME) and the

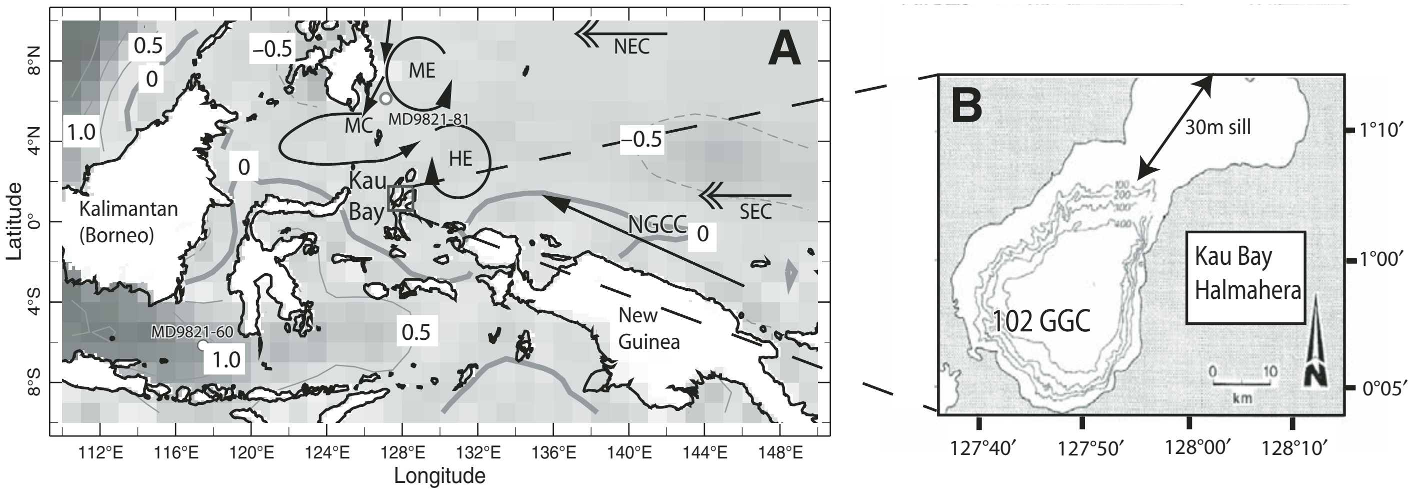

Figure 1. A: Location of Kau Bay in western tropical Pacific. Contours and grayscale shading are sea surface temperature anomalies during peak of very strong 1997–1998 El Niño event (Reynolds and Smith, 1994). Arrows represent general direction of ocean currents discussed in text. NEC—Northern Equatorial Current; SEC—Southern Equatorial Current; NGCC—New Guinea Coastal Current; MC—Mindanao Current; ME— Mindanao Eddy; HE—Halmahera Eddy. Locations of sediment cores analyzed by Newton et al. (2006) (MD9821–60) and Stott et al. (2004) (MD9821–81) are shown (see text). B: Kau Bay bathymetry with contours from 100 m to 400 m.

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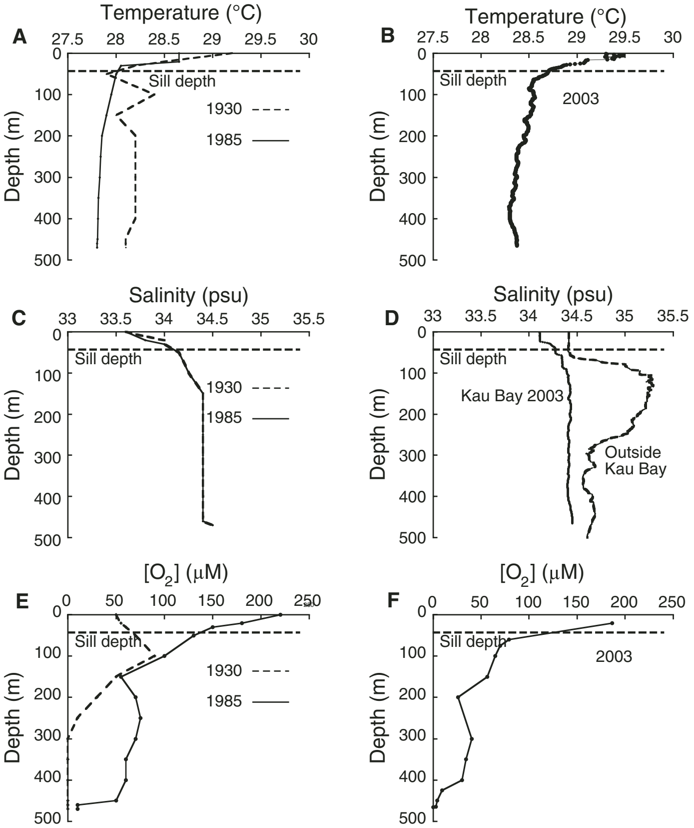

Figure 2. Water-column profiles of temperature (A and B), salinity (C and D), and oxygen (E and F) from 1930 (Van Riel, 1943), 1985 (Van der Weijden et al., 1989), and measurements from 2003.

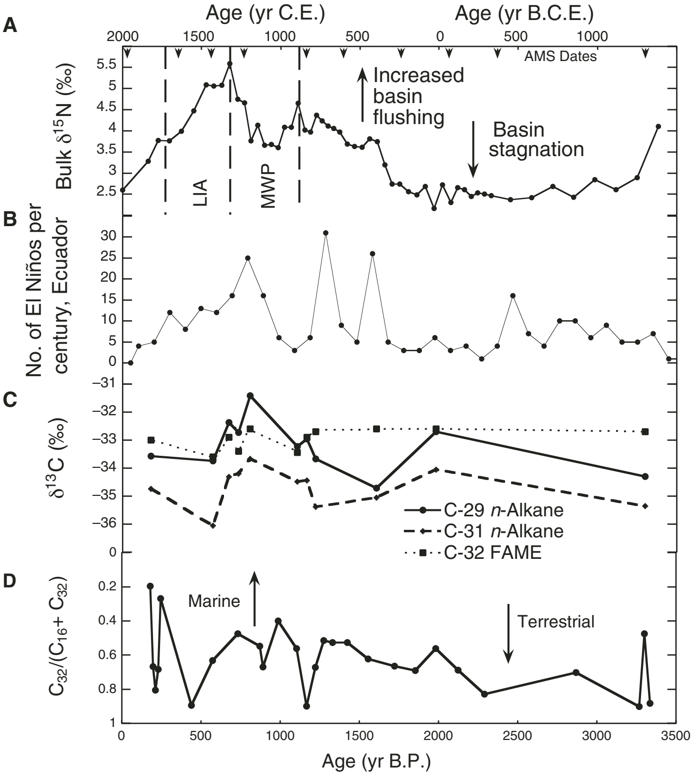

Figure 3. A: Bulk sediment δ^{15}N. LIA—Little Ice Age; MWP—Medieval Warm Period. B: Total number of strong El Niño events per century interpreted from flood deposits in Laguna Pallcacocha, Ecuador (Moy et al., 2002). C: δ^{13}C of terrestrial plant waxes (*n*-alkanes with chain lengths of 29 and 31 carbons, fatty acids with chain lengths of 32 carbons). D: Ratio of terrestrial to marine fatty acids.

anticyclonic Halmahera Eddy (HE). The strength of these eddies depends upon the strength of the MC and NGCC, so their presence may influence Kau Bay flushing (Arruda and Nof, 2003; Masumoto et al., 2001; Ueki et al., 2003; Wyrtki, 1961) (Fig. 1).

There is also evidence that interannual changes in the mean climate state of the western equatorial Pacific, related to El Niño Southern Oscillation (ENSO) activity, exert significant control on the ventilation of Kau Bay. During modern El Niño events, the mixed layer around Halmahera is characterized by colder and saltier water. Sea surface temperature data (Reynolds and Smith, 1994) for the 1° × 1° grid near the Kau Bay entrance reveal that sea surface temperature cooled ~1 °C during all El Niño events since 1981. Mooring data collected at long 138°E and 142°E (Ueki et al., 2003) show that the typical seasonal variability in the flow direction of the NGCC ceased during the 1997–1998 El Niño and that instead the NGCC flowed northwestward all year, advecting cold and salty surface water toward Halmahera. Moreover, precipitation in the area of Halmahera is not significantly affected by the Asian Monsoon, but is strongly influenced by ENSO, with lower precipitation during El Niño events (Aldrian and Dwi Susanto, 2003). Increased primary productivity around Halmahera and Kau Bay during the very strong El Niño of 1997–1998 (Christian et al., 2004) is consistent with the proposed thermocline shoaling associated with El Niño–driven changes in regional circulation (Arruda and Nof, 2003; Ueki et al., 2003).

SEDIMENTARY NITROGEN AND CARBON ISOTOPES

We measured nitrogen isotope ratios (as δ^{15}N) on the <63 μm fraction of bulk sediment from Kau Bay gravity core BJ8–03–102GGC at 8 cm intervals (4.25 m long, 377 m water depth; Fig. 1). Radiocarbon dates on pteropods indicate that the δ^{15}N series has a resolution of ~1 sample per 60 yr (Fig. 3; see methods, Table DR1 in the GSA Data Repository[1]). Over the past 3500 yr, sedimentary δ^{15}N varied between 2.2‰ and 5.6‰. Today, nitrate (NO_3^-) is completely consumed in Kau Bay's surface water (Fig. 4) so that sedimentary δ^{15}N records changes in the isotopic composition of subeuphotic zone nitrate. Several processes likely contribute to the δ^{15}N of nitrate in Kau Bay: (1) inputs from the open ocean; (2) inputs by N fixation; (3) removal via denitrification in anoxic water; and (4) nitrification of ammonium fluxing out of the anoxic water and sediment. Western Pacific surface water NO_3^- has a δ^{15}N of ~5‰–6‰ (samples collected during the R/V *Baruna Jaya VIII* 2003 cruise). Nitrogen fixation reduces surface water NO_3^- δ^{15}N because nitrogen with a δ^{15}N of 0‰ is fixed from atmospheric N_2. In the deep basin, water-column denitrifying bacteria in the absence of significant oxygen generally enrich the water-column nitrate pool in $^{15}NO_3^-$ through the preferential conversion of $^{14}NO_3^-$ to N_2 and N_2O gases (Brandes et al., 1998; Liu and Kaplan, 1989; Sigman et al., 2003, 2005). In contrast, nitrification, the oxidation of ammonium to nitrate, via nitrite, has a relatively large negative fractionation (~15‰; Casciotti et al., 2002) that is rarely apparent in oxic water columns because of the short turnover time for ammonium. Ammonium builds up in high concentrations in anoxic sediment, which may become a steady source of ammonium to the overlying oxic water column. The enrichment in Kau Bay bottom water δ^{15}N (Fig. 4) is not as high as expected from the observed enrichment in bottom water δ^{18}O of NO_3. In culture they increase 1:1

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[1] GSA Data Repository item 2008205, methods and Table DR1, is available online at www.geosociety.org/pubs/ft2008.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

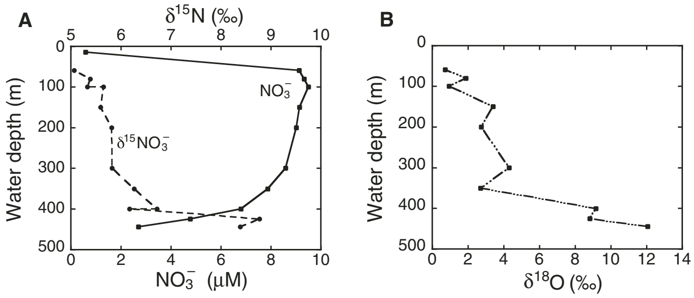

Figure 4. Nitrate concentrations and δ^{15}N (A) and δ^{18}O (B) of nitrate from center of Kau Bay, station 113HC; measurements taken in July 2003.

(Granger et al., 2004a, 2004b). In Kau Bay δ^{18}O of NO$_3^-$ increases from 2‰ to 12‰ between 350 m and the bottom, while δ^{15}N increases from 6‰ to 9‰. This suggests that ~7‰ of the enrichment expected from the δ^{18}O is negated by input of isotopically depleted N. The deviation in the expected relationship appears to be originating from the sediment-water interface, and we infer nitrification of NH$_4$ to be the source.

In open ocean regions of denitrification such as the eastern tropical Pacific and Arabian Sea, sedimentary δ^{15}N is relatively high due to the effects of incomplete denitrification in the oxygen minimum zone (Liu et al., 1989; Altabet, 2001; Ganeshram et al., 2000). In contrast, nitrate supply to semi-enclosed basins such as Kau Bay is limited, and denitrification results in the near complete removal of nitrate at depth. Moreover, extensive denitrification reduces the N/P in the water column, creating ideal conditions for nitrogen fixers (Haug et al., 1998; Thunell et al., 2004). Ultimately, an increase in denitrification enhances nitrogen fixation, which results in lower δ^{15}N, and vice versa (Haug et al., 1998; Thunell et al., 2004; Deutsch et al., 2007). In Kau Bay, nearly all the nitrate below the oxycline is consumed, and the proportion of denitrified nitrate bearing the ^{15}NO$_3^-$-rich signature of denitrification is low relative to the overlying surface water nitrate pool (Fig. 4). An isotope effect (ε) of ~1.5‰ is calculated in Kau Bay, assuming a closed system based on the Rayleigh approximation $\delta^{15}\text{N} = \delta^{15}\text{N}_{\text{mid-water column}} - \varepsilon \ln f$ (Mariotti et al., 1981; Altabet and Francois, 1994), where the fraction of unused nitrate, f, is $(\text{NO}^-_{3\ \text{sediment water interface}})/(\text{NO}^-_{3\ \text{mid-water column}})$ (~10 μM). There is minimal expression of the denitrification ε of ~20‰–30‰ (Thunell et al., 2004; Sigman et al., 2003) in Kau Bay.

One interpretation of the downcore sedimentary δ^{15}N data in Kau Bay is that the isotopic composition of the surface nitrate pool reflects the combined effects of this nitrogen fixation-denitrification feedback and inputs from the open ocean (~5‰) (Haug et al., 1998; Thunell et al., 2004). Alternatively, the downcore intervals of lower sedimentary δ^{15}N may reflect periods of enhanced terrestrial inputs to Kau Bay (terrestrial organic matter has an average δ^{15}N of ~0‰ (Brandes and Devol, 2002) or enhanced inputs of isotopically light N via the nitrification pathway.

The −21‰ to −22‰ δ^{13}C values of sedimentary organic carbon (Table 1) indicate that marine organic carbon dominates throughout the core. The concentrations and relative abundance of terrestrial to marine fatty acids (Fig. 3D) suggest a gradual decrease in the influence of a terrestrial source for organic matter in Kau Bay concomitant with the increase in δ^{15}N from ca. 1600 yr B.P. to 700 yr B.P. The consistently low δ^{13}C values (from −36‰ to −31‰) of terrestrial *n*-alkanes and fatty acids clearly indicate that inputs from C$_3$ land plants (Makou et al., 2007; Street-Perrott et al., 1997) did not exert the dominant control on bulk sedimentary organic δ^{13}C values (Table 1) throughout the past 3500 yr.

Variations in the δ^{15}N and δ^{13}C therefore appear to primarily reflect water-column processes (Street-Perrott et al., 1997). Moreover, whether the variations in δ^{15}N are related to enhanced N fixation or enhanced nitrification, low δ^{15}N corresponds to intervals of intensified stratification and anoxia in the basin and higher δ^{15}N reflects periods of increased ventilation and the input of open ocean nitrate.

TABLE 1. PERCENT C (ORGANIC), PERCENT N (TOTAL), C/N, AND δ^{13}C ORGANICS FROM 102 GGC, KAU BAY, INDONESIA

Age (yr B.P.)	Organic C (%)	Total N (%)	C$_{\text{org}}$/N$_{\text{tot}}$	δ^{13}C (‰)
163	4.58	0.46	11.68	–21.75
578	3.83	0.34	13.19	–22.06
1035	4.24	0.35	13.97	–22.32
1326	3.78	0.32	13.89	–21.66
2329	4.53	0.35	15.28	–22.19
3285	3.89	0.29	15.59	–21.27

The core top (modern) sample from Kau Bay multicore 103MC-F (companion to BJ8–03–102GGC) has a δ^{15}N of 2.6‰ while subeuphotic zone nitrate has a δ^{15}N of ~5.5‰ (Figs. 3A and 4). The apparent disconnect between the modern water-column nitrate pool and surface sediment δ^{15}N may be accomplished through intensified terrestrial inputs or a recent partial flushing of the water column (Fig. 4). Recent ventilation is consistent with the long La Niña phase from 1998 to 2002 and a transition to more El Niño conditions in 2003. If this is the cause, then this offset highlights the extremely dynamic nature of the Kau Bay water column.

DISCUSSION

We interpret downcore increases in the δ^{15}N as reflecting enhanced Kau Bay ventilation. Although we cannot unequivocally ascertain which of the proposed processes is responsible for the observed changes in sedimentary δ^{15}N, it is important to note that whether acting alone or together, all generate the same response to ventilation; increased ventilation will lead to higher δ^{15}N while stagnation will reduce the δ^{15}N. Increased flushing is most likely stimulated during periods of more frequent and/or intense El Niño events or a more El Niño–like mean state in the WPWP. A reduction in El Niño frequency and/or intensity or fresher and warmer mean state in the WPWP would result in basin stagnation and an overall decrease in δ^{15}N. Accordingly, the δ^{15}N record (Fig. 3A) documents a less El Niño–like (neutral or La Niña–like) mean state or less frequent and/or weaker El Niño episodes from ca. 3500 to ca. 1700 yr B.P. During this time interval, high runoff likely caused the increase in terrestrial input and may have promoted a freshwater cap at the basin's surface that resulted in basin stagnation. Ventilation improved ca. 1700 yr B.P., likely due to thermocline shoaling in the WPWP in association with more El Niño–like mean surface conditions or stronger and/or more frequent El Niño events. Basin stagnation, signaling less El Niño–like conditions, occurred during the time frame of the Medieval Warm Period (MWP), from ca. 1000 to 750 yr B.P. This episode was followed by an increase in El Niño activity that culminated at the beginning of the Little Ice Age ca. 700 yr B.P. The Kau Bay record suggests that the remainder of the Little Ice Age was characterized by a steady decrease in El Niño activity with warming and freshening of the surface water that continued to the present. The surface freshening is consistent with the results of Stott et al. (2004) and Newton et al. (2006).

Within age model uncertainties, other paleoclimatic records support our interpretation of the Kau Bay geochemical records as reflecting century-scale ENSO variability. Most notably, the chronology of flood deposits in Laguna Pallcacocha, Ecuador (Moy et al., 2002; Rodbell et al., 1999), attributed to intense El Niño events, shows similar century-scale periods of increased El Niño frequency over the past ~1500 yr, with diminished El Niño frequency during the past ~700 yr (Fig. 3). Decreased terrestrial input on the Peru margin and in the Cariaco Basin that began ca. 1000 yr B.P. has been attributed to drought and is also consistent with less frequent or weaker El Niño events or less El Niño–like conditions (Haug et al., 2001; Hodell et al., 2005; Rein et al., 2004, 2005). Not all climatic events recorded in the Kau Bay and Laguna Pallcacocha are evident in these other marine records, suggesting that they may be influenced by other climatic factors. By contrast, the finding of similar century-scale variability in climate archives from two El Niño–sensitive regions on opposite sides of the tropical Pacific strongly suggests that they are dominated by the low-frequency variability of ENSO or by ENSO-related changes in the mean state of the surface ocean in equatorial Pacific.

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