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Climate and hydrographic variability in the Indo-Pacific Warm Pool during the last millennium

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[1] Planktonic foraminiferal Mg/Ca and $\delta^{18}O$ derived sea surface temperature and salinity records from the Makassar Strait, Indonesia, show a long-term cooling and freshening trend, as well as considerable centennial-scale variability during the last millennium. The warmest temperatures and highest salinities occurred during the Medieval Warm Period (MWP), while the coolest temperatures and lowest salinities occurred during the Little Ice Age (LIA). These changes in the western Pacific, along with observations from other high resolution records indicate a regionally coherent southern displacement of the Inter-tropical Convergence Zone during the LIA, with more arid conditions in the northern tropics and wetter conditions in the southern tropics. **Citation:** Newton, A., R. Thunell, and L. Stott (2006), Climate and hydrographic variability in the Indo-Pacific Warm Pool during the last millennium, *Geophys. Res. Lett.*, *33*, L19710, doi:10.1029/2006GL027234.

1. Introduction

[2] The Indo-Pacific Warm Pool (IPWP) is one of the warmest regions in the modern oceans, and consequently, air-sea interactions in this region strongly influence global heat and water vapor exchange between the ocean and atmosphere. Specifically, changes in sea surface temperatures and convection in the tropical Indo-Pacific region are responsible for much of the interannual (i.e. ENSO) to decadal (i.e. PDO) climate variability observed in extra-tropical regions [*Cane*, 1998; *Hoerling et al.*, 2001; *Sun et al.*, 2003]. *Cane and Clement* [1999] used coupled ocean-atmosphere models to demonstrate that heat and water vapor flux from the tropics may also be important in affecting global climate change on a variety of different time scales. Although past changes in the IPWP have been described on millennial and longer timescales [*Lea et al.*, 2000; *Stott et al.*, 2002, 2004; *Visser et al.*, 2003; *de Garidel-Thoron et al.*, 2005; *Holbourn et al.*, 2005], few studies have examined decadal to centennial–scale variability in this climatically important region.

[3] The last millennium has been marked by several global-scale climate fluctuations. In particular, the Little Ice Age (LIA; ~1400–1850 AD) was one of the largest amplitude events recorded in many regions since the Last Glacial Maximum [*Lamb*, 1982; *Bradley et al.*, 2003]. LIA climatic changes have been reconstructed from a number of high latitude locations. In the Northern Hemisphere, pronounced cold periods occurred across continental North America and Eurasia during the seventeenth and nineteenth centuries [*Mann et al.*, 1998], with a general cooling trend present from the fifteenth century to the end of the nineteenth century [*Mann et al.*, 1999]. Ice cores from both Greenland and Antarctica show a nearly synchronous onset of cooling [*Kreutz et al.*, 1997], while marine sediment records from North Atlantic record increased sea ice drift during this period [*Bond et al.*, 1999, 2001]. Numerous other studies of marine cores, tree rings, and historical records also show mid- to high latitude temperature decreases and glacial advances during the LIA [*Broecker*, 2001].

[4] The extent and severity of the LIA is less constrained in tropical regions. For example, *Hendy et al.* [2002] assumed that the LIA cooling was restricted to higher latitudes and thus resulted in an increased latitudinal SST gradient and enhanced poleward transfer of heat during this time period. Using corals, *Gagan et al.* [2000] found a distinct cool interval from 1800 to 1840 in the tropical Indian and Pacific Oceans, while *Cobb et al.* [2003] found evidence of increased El Niño activity during the seventeenth century. Whereas corals are excellent proxies for reconstructing high resolution climate variability, they also tend to provide only relatively short records that span several hundred years or less. By analyzing planktonic foraminifera from a continuous marine core taken from a high sedimentation site we are able to generate high resolution sea surface temperature (SST) and salinity (SSS) records over a longer time interval. Here we present ~1,000 year-long records of SST and SSS from the Makassar Strait, Indonesia derived from paired Mg/Ca and $\delta^{18}O$ measurements of planktonic foraminifera.

2. Core Material and Methods

[5] The core used for this study, MD9821-60, was collected aboard the *Marion Dufrense* in 1998 as part of the IMAGES coring program. It was collected at 5°12.07 S, 117°29.20 E from a water depth of 1185 m. This depth is well above the present-day lysocline [*Farrell and Prell*, 1989], which allows for excellent carbonate preservation. Although the uppermost sediments representing roughly the last 150 years were lost during the coring process, the average Holocene sedimentation rate at this location is well over 100 cm per 1,000 years, making it an ideal core for high resolution studies of hydrographic changes in the Makassar Strait.

[6] The age model for the core is based on linear extrapolation between three radiocarbon dates and a 2 cm

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Table 1. AMS Results for Mixed G. Sacculifer and G. Ruber Samples[a]

Depth, cm	^{14}C Age	Std. Dev.	Calibrated YBP[b]	1σ Range (YBP)[b]	Calendar Age (AD)[b]
15 (tephra)					1815
36	720	±60	280	174–382	1670
72	1005	±35	520	479–597	1430
138	1565	±40	1040	947–1115	910

[a]Each sample weighed between 1.5 and 3 mg.
[b]All ages were calibrated using the CALIB 4.1 program [*Stuiver and Reimer*, 1993]. A standard reservoir age correction (ΔR) of 75 ± 80 years for the Indian Ocean and southeast Asia was used [*Southon et al.*, 2002].

thick ash layer. This is the only visibly distinctive ash layer in the upper portion of the core and it is assumed to be from the 1815 Tambora eruption [*Stothers*, 1984]. All ^{14}C measurements were made on mixed samples of *Globigerinoides sacculifer* and *Globigerinoides ruber* and performed at the Center for Accelerator Mass Spectrometry at Lawrence Livermore National Laboratory (Table 1). A depth vs. age plot shows a fairly continuous rate of sedimentation over the last 1,000 years with some stretching in the uppermost 50 cm due to the core recovery process.

[7] The core was sampled continuously at 1 cm intervals, providing a time resolution of less than 10 years. Paired Mg/Ca and δ^{18}O analyses of *G. ruber* were carried out on all samples in order to generate records of past SST and SSS. For the trace metal analyses, each sample underwent a rigorous cleaning method using a procedure modified from *Boyle* [1981] to eliminate possible contamination from organic matter and silicates. Magnesium and calcium were simultaneously measured using a Jobin Yvon Ultima Inductively Coupled Plasma Atomic Emission Spectrophotometer (ICP-AES). The 1σ analytical precision in this study based on analyses of standards is better than 1.0%. The stable isotope analyses were carried out using a VG Optima stable isotope ratio mass spectrometer (IRMS) equipped with an automated carbonate system. The δ^{18}O data are reported in delta notation relative to the (Vienna) Pee Dee Belemnite (VPDB) standard. The long-term standard reproducibility for oxygen (δ^{18}O) isotopes based on replicate measurements of our reference standard is ±0.07‰.

[8] In the equatorial western Pacific, the production of *G. ruber* is highest during boreal summer [*Kawahata et al.*, 2002] and thus sediment assemblages of this species should be biased or weighted towards this season. The δ^{18}O of foraminiferal calcite (δ_c) reflects the combined effects of the temperature of calcification (T) and the local oxygen isotopic of sea water (δ_w), which varies as a function of salinity, whereas the Mg/Ca ratio of these shells is primarily temperature dependent [*Lea et al.*, 1999]. The Mg/Ca-temperature relationship for *G. ruber* determined by *Dekens et al.* [2002] was used in this study. The standard error of estimate for this equation is 1.2°C. By measuring both δ_c and Mg/Ca on the same samples, it is possible to isolate the salinity component in the δ^{18}O signal. Using the measured δ_c values and the calculated Mg/Ca temperatures (T) we can estimate δ_w using the *Bemis et al.* [1998] paleotemperature equation for *G. ruber*

$$T(^\circ C) = 14.9 - 4.80(\delta_c - \delta_w). \quad (1)$$

The standard error for this equation is 0.5°C. δ_w was converted to salinity using the modern salinity- δ_w relationship for this region [*Morimoto et al.*, 2002]

$$\delta w(SMOW) = 0.42S - 14.3. \quad (2)$$

3. Temperature and Salinity Changes

[9] Using these procedures we have generated records of climate and hydrographic variability in the Makassar Strait region extending back to ~1000 AD that are characterized both by centennial and decadal-scale variability (Figure 1). The SST and SSS estimates for the southern Makassar Strait primarily range from 28 to 30°C and 33.5 to 36.0, respectively, over the entire time period. The average modern summer SST and SSS are 29°C and 35, respectively. The period from ~1000 to 1400 AD is marked by the warmest temperatures and highest salinities. During this time interval, the average SSTs were between 29–30°C (mean of 29.4°C) and salinities fluctuated between ~34–35.8 (mean of 34.9), and are very similar to present day conditions in this area. This 400-year long period of warm temperatures and high salinities is equivalent in age to the Medieval Warm Period (MWP), a time when radiative forcing was high [*Lean et al.*, 1995]. The MWP has been well documented in mid to high latitude temperature records from the northern hemisphere [*Mann et al.*, 1999]. In the tropics, arid conditions existed in northern South America [*Haug et al.*, 2001] and eastern Africa [*Verschuren et al.*, 2000] during the MWP, although central Pacific corals showed no indication of a warming at this time [*Cobb et al.*, 2003].

[10] For our study site, there is a significant long term cooling and freshening trend during the last millennium. Mean temperature decreased by approximately 1°C during the last 1,000 years, while salinity decreased by 0.9. This is consistent with an overall trend of decreasing salinity and temperature in the IPWP throughout the Holocene [*Stott et al.*, 2004]. Specifically, the SST record shows a distinct cooling trend beginning at ~1400 AD and lasting for several hundred years, a period equivalent in time to the Little Ice Age (LIA). In particular, the lowest temperatures (~28°C) occur around 1700 A.D., during the period of reduced solar intensity known as the Maunder Minimum. Since the production of *G. ruber* in this region of the tropics is highest during the summer [*Kawahata et al.*, 2002], these estimated temperatures are indicative of summer SSTs, which were 1.0–1.5°C cooler than present. The magnitude of this LIA cooling is approximately half of that reported for the last glacial maximum in the IPWP [*Lea et al.*, 2000;

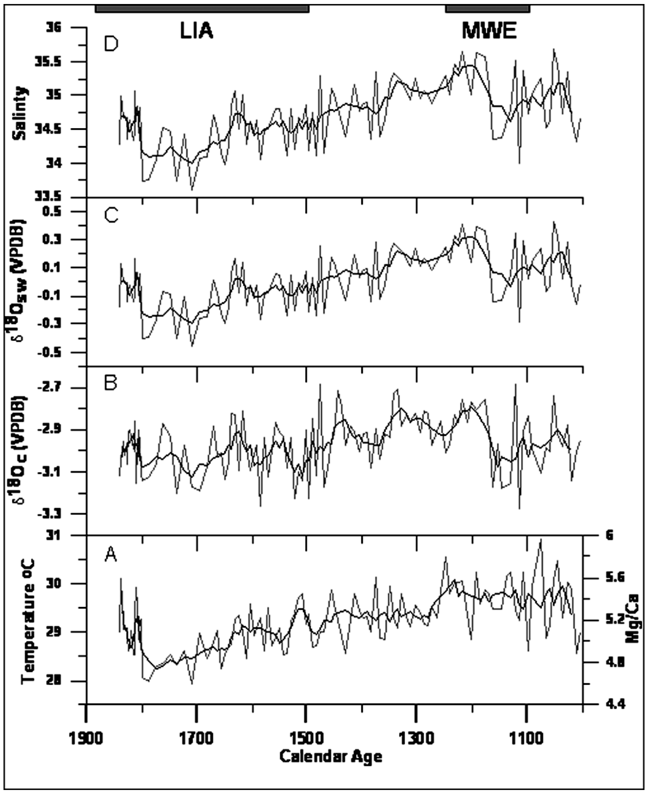

Figure 1. (a) *Globigerinoides ruber* Mg/Ca (right axis) and derived temperature estimates. (b) Measured δ^{18}O of calcite. (c) Mg/Ca and δ^{18}O derived seawater δ^{18}O. D. Reconstructed sea surface salinity. The bars denote the Little Ice Age (LIA) and the Medieval Warm Epoch (MWE), the period of increased solar radiation that occurred during the longer Medieval Warm Period.

Visser et al., 2003]. These results clearly indicate a climatic cooling during the LIA that extended well outside the higher northern latitudes. In fact, the recognition that that tropical Pacific warm pool temperatures were as much as 1.5°C cooler during the LIA must be considered an important factor itself in establishing what caused the climate to cool as it did.

4. Tropical Climate During the LIA

[11] Our findings, when integrated with other high resolution climate records from the tropics and subtropics provide a regionally variable but coherent picture of hydrological conditions in the low latitudes during the LIA (Table 2 and Figure 2). Titanium concentrations in Cariaco Basin (~10°N) sediments suggest increased aridity in this sector of the tropical Atlantic during the LIA [*Haug et al.*, 2001], while paired coral δ^{18}O and Mg/Ca records from the Caribbean Sea (19°N) show that the LIA was marked by a 2°C SST decrease and higher surface salinities [*Wantanabe et al.*, 2001]. Ostracod δ^{18}O from lacustrine sediments from the Yucatan Peninsula (20°N) suggest arid conditions during the LIA [*Hodell et al.*, 2005], while paired foraminiferal δ^{18}O and Mg/Ca records from the Dry Tortugas also suggest decreased precipitation at this time [*Lund and Curry*, 2006]. Similarly, weaker monsoons and more arid conditions are reported for the LIA from studies of marine sediments from the Arabian [*Anderson et al.*, 2002] and South China seas [*Wang et al.*, 1999], and tree rings from Pakistan [*Treydte et al.*, 2006]. A similar decrease in precipitation during the LIA is evident in coral records from the Gulf of Chiriqui (~7°N) in the eastern Pacific [*Linsley et al.*, 1994]. Not only do all six of these studies indicate more arid conditions during the LIA (Table 2) but all of the study sites are located close to the present day position of the ITCZ during boreal summer (July) (Figure 2). A different scenario occurs south of the equator. Records from Lake Titicaca (~15°S) in South America show enhanced precipitation during the LIA [*Baker et al.*, 2001]. In east Africa records from Lake Malawi (~10°S) show periods of enhanced diatom productivity during the LIA, most likely a result of enhanced northerly winds associated with southern excursions of the ITCZ [*Brown and Johnson*, 2005]. Lake-level fluctuations in Lake Naivasha, Kenya (0.5°S) show maxima occurring during the LIA indicating increased precipitation [*Verschuren et al.*, 2000]. These southern hemisphere study sites are all in proximity to the present day boreal winter position of the ITCZ (Figure 2). Overall, the trends seen in these Indo-Pacific and Atlantic records may represent a continuation of the Holocene southward migration of the Inter-tropical Convergence Zone (ITCZ) into the last millennium [*Haug et al.*, 2001]. Specifically, the development of more arid conditions in the northern tropics/subtropics and wetter conditions south of the equator during the LIA is indicative of a pronounced and rapid southward displacement of the ITCZ and associated band of high precipitation during this time period.

[12] In the modern ocean, the persistent northern bias of the ITCZ arises as a result of the cool upwelled water dominating the temperature signal of the eastern equatorial Pacific (EEP)

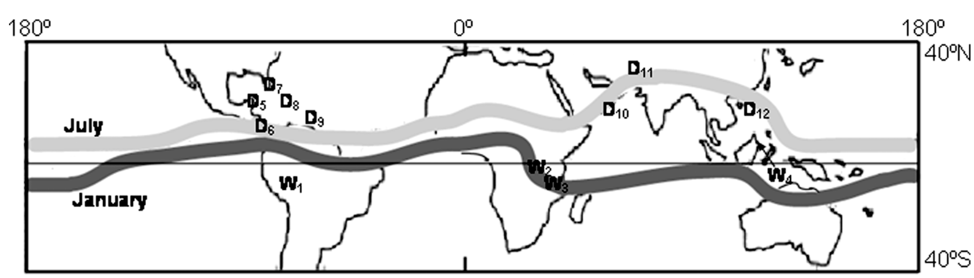

Figure 2. The map is showing the modern locations of the ITCZ in July and January. Each location that was drier during the Little Ice Age is marked with a D. Locations that were wetter are marked with a W. Further information about the evidence at each location is found in Table 2.

Table 2. Listing of Each Locality and the Conditions Observed at That Site During the Little Ice Age[a]

	Location	LIA Conditions Relative to Modern	Data Source
1	Lake Titicaca, Peru	Wetter - high lake levels	*Baker et al.* [2001]
2	Lake Naivasha, Kenya	Wetter – high lake levels	*Verschuren et al.* [2000]
3	Lake Malawi, Malawi	Increased productivity and wind strength	*Brown and Johnson* [2005]
4	Makassar Straits, Indonesia	Wetter – lower surface salinities	This paper
5	Yucatan Peninsula	Drier – increased lake salinity	*Hodell et al.* [2005]
6	Gulf of Chiriqui, Panama	Drier – higher surface salinities	*Linsley et al.* [1994]
7	Dry Tortugas, Florida Current	Drier – higher sea surface salinities	*Lund and Curry* [2006]
8	Puerto Rico, Caribbean Sea	Drier – increased sea surface salinity	*Wantanabe et al.* [2001]
9	Cariaco Basin, Venezuela	Drier – decreased river input	*Haug et al.* [2001]
10	Oman Margin, Arabian Sea	Drier – weaker summer monsoon	*Anderson et al.* [2002]
11	Karakorum Range Vallies, Pakistan	Drier – decreased precipitation	*Treydte et al.* [2006]
12	South China Sea	Drier – higher surface salinities	*Wang et al.* [1999]

[a]The number on the far left refers to the position in Figure 2.

[*Haug et al.*, 2001]. This results in a displaced SST maximum into the Northern Hemisphere. However, during the LIA, probable El-Niño related warming resulted in reduced upwelling conditions, leading to decreased cooling in the EEP [*Cobb et al.*, 2003]. These changes could have reduced the differential heating between the Northern and Southern hemispheres, and resulted in a southward migration of the ITCZ. Such a scenario would explain the increase in aridity north of the equator and enhanced precipitation in the southern tropics [*Baker et al.*, 2001].

5. Conclusions

[13] Our records from the Makassar Strait show that climate changes during the Medieval Warm Period and Little Ice Age were not confined to the high latitudes. Particularly, during the LIA, Makassar Strait SST were $\sim$1.5° cooler and SSS were $\sim$1 lower. This indicates enhanced precipitation and/or decreased evaporation during this time in the western equatorial Pacific. Marine and terrestrial records indicate increased precipitation during the LIA in southern tropical South America and southern tropical East Africa as well. Conversely, a variety of records from Central America, the Caribbean, and the northern Indian Ocean suggest the LIA was marked by periods of increased aridity. This apparent latitudinal difference in precipitation is attributed to a rapid southward migration of the ITCZ. In the modern Atlantic, the ITCZ is centered over Central and northernmost South America in July and over equatorial South America in January. A southern displacement of the boreal summer ITCZ during the LIA to a position similar to the modern January position would direct precipitation away from the Caribbean and toward equatorial and southern tropical South America.

[14] In the Indo-Pacific, a southward displacement of the ITCZ toward the modern January location would bring additional precipitation to the Makassar Strait and East African lake region, while precipitation north of the equator would decrease.

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References

Anderson, D. M., J. T. Overpeck, and A. K. Gupta (2002), Increase in the Asian southwest monsoon during the past four centuries, *Science*, *297*, 596–599.

Baker, P. A., G. O. Seltzer, S. L. Fritz, R. B. Dunbar, M. J. Grove, P. M. Tapia, S. L. Cross, H. D. Rowe, and J. P. Broda (2001), The history of South America tropical precipitation for the past 25,000 years, *Science*, *291*, 640–643.

Bemis, B. E., H. J. Spero, J. Bijima, and D. W. Lea (1998), Reevaluation of the oxygen isotopic composition of planktonic foraminifera: Experimental results and revised paleotemperature equations, *Paleoceanography*, *13*, 150–160.

Bond, G. C., W. Showers, M. Elliot, M. Evans, R. Lotti, I. Hajdas, G. Bonani, and S. Johnson (1999), The North Atlantic's 1–2 kyr climate rhythm: Relation to Heinrich events, Dansgaard/Oeschger cycles, and the Little Ice Age, in *Mechanisms of Global Climate Change at Millennial Timescales*, *Geophys. Monogr. Ser.*, vol. 122, edited by P. Clark, R. Webb, and L. Keigwin, pp. 35–58, AGU, Washington, D. C.

Bond, G. C., et al. (2001), Persistent solar influence on North Atlantic climate during the Holocene, *Science*, *294*, 2130–2136.

Boyle, E. A. (1981), Cadmium, copper, zinc, and barium in foraminiferal tests, *Earth, Planet. Sci. Lett.*, *53*, 11–35.

Bradley, R. S., K. R. Briffa, J. Cole, M. K. Hughes, and T. J. Osborn (2003), The climate of the last millennium, in *Paleoclimate, Global Change, and the Future*, edited by K. Alverson, R. S. Bradley, and T. F. Pederson, pp. 105–141, Springer, New York.

Broecker, W. (2001), Was the Medieval Warm Period global?, *Science*, *291*, 1497–1499.

Brown, E. T., and T. C. Johnson (2005), Coherence between tropical East African and South American records of the Little Ice Age, *Geochem. Geophys. Geosyst.*, *6*, Q12005, doi:10.1029/2005GC000959.

Cane, M. A. (1998), A role for the tropical Pacific, *Science*, *282*, 59–60.

Cane, M. A., and A. Clement (1999), A role for the tropical Pacific coupled ocean-atmosphere system on Milankovitch and millennial timescales part II: Global impacts, in *Mechanisms of Global Climate Change at Millennial Timescales*, *Geophys Monogr. Ser.*, vol. 112, edited by P. Clark, R. Webb, and L. Keigwin, pp. 373–383, AGU, Washington, D. C.

Cobb, K. M., C. D. Charles, H. Cheng, and R. L. Edwards (2003), El Niño/Southern Oscillation and tropical Pacific climate during the last millennium, *Nature*, *424*, 271–276.

de Garidel-Thoron, T., Y. Rosenthal, F. Bassinot, and L. Beaufort (2005), Stable sea surface temperatures in the western Pacific warm pool over the past 1.75 million years, *Nature*, *433*, 294–298.

Dekens, P. S., D. W. Lea, D. K. Pak, and H. J. Spero (2002), Core top calibration of Mg/Ca in tropical foraminifera: Refining paleotemperature estimation, *Geochem. Geophys. Geosyst.*, *3*(4), 1022, doi:10.1029/2001GC000200.

Farrell, J. W., and W. L. Prell (1989), Climatic change and $CaCO_3$ preservation: An 800,000 year bathymetric reconstruction from the central Pacific Ocean, *Paleoceanography*, *4*, 447–466.

Gagan, M. K., L. K. Ayliffe, J. W. Beck, J. E. Cole, E. R. M. Druffel, R. B. Dunbar, and D. P. Schrag (2000), New views of tropical paleoclimates from corals, *Quat. Sci. Rev.*, *19*, 45–64.

Haug, G. H., K. A. Hughen, D. M. Sigman, L. C. Peterson, and U. Rohl (2001), Southward migration of the intertropical convergence zone through the Holocene, *Science*, *293*, 1304–1308.

Hendy, E. J., M. K. Gagan, C. A. Alibert, M. T. McCulloch, J. M. Lough, and P. J. Isdale (2002), Abrupt decrease in tropical Pacific sea surface salinity at the end of Little Ice Age, *Science*, *295*, 1511–1514.

Hodell, D. A., M. Brenner, J. H. Curtis, R. Mendina-Gonzalez, E. I. C. Can, A. Albornaz-Pat, and T. P. Guilderson (2005), Climate change on the Yucatan Peninsula during the Little Ice Age, *Quat. Res.*, *63*, 109–121.

Hoerling, M. P., J. W. Hurrell, and T. Xu (2001), Tropical origins for recent North Atlantic climate change, *Science*, *292*, 90–92.

Holbourn, A., W. Kuhnt, H. Kawamura, Z. Jian, P. Grootes, H. Erlenkeuser, and J. Xu (2005), Orbitally paced paleoproductivity variations in the Timor Sea and Indonesian Throughflow variability during the last 460 kyr, *Paleoceanography*, *20*, PA3002, doi:10.1029/2004PA001094.

Kawahata, H., A. Nishimura, and M. K. Gagan (2002), Seasonal change in foraminiferal production in the western equatorial Pacific warm pool: Evidence from sediment trap experiments, *Deep Sea Res., Part II*, *49*, 2783–2800.

Kreutz, K. J., P. A. Mayewski, L. D. Meeker, M. S. Twickler, S. I. Whitlow, and I. I. Pittalwala (1997), Bipolar changes in atmospheric circulation during the Little Ice Age, *Science*, *277*, 1294–1296.

Lamb, H. H. (1982), *Climate, History, and the Modern World*, Routledge, Boca Raton, Fla.

Lea, D. W., T. A. Mashiotta, and H. J. Spero (1999), Controls on magnesium and strontium uptake in planktonic foraminifera determined by live culturing, *Geochim. Cosmochim. Acta*, *63*, 2369–2379.

Lea, D. W., D. K. Pak, and H. J. Spero (2000), Climate impact of late Quaternary equatorial Pacific sea surface temperature variations, *Science*, *289*, 1719–1724.

Lean, J., J. Beer, and R. Bradley (1995), Reconstruction of solar irradiance since 1610: Implications for climate change, *Geophys. Res. Lett.*, *22*, 3195–3198.

Linsley, B. K., R. B. Dunbar, G. M. Wellington, and D. A. Mucciarone (1994), A coral-based reconstruction of intertropical convergence zone variability over Central America since 1707, *J. Geophys. Res.*, *99*, 9977–9994.

Lund, D. C., and W. Curry (2006), Florida Current surface temperature and salinity variability during the last millennium, *Paleoceanography*, *21*, PA2009, doi:10.1029/2005PA001218.

Mann, M. E., R. S. Bradley, and M. K. Hughes (1998), Global-scale temperature patterns and climate forcing over the past six centuries, *Nature*, *392*, 779–787.

Mann, M. E., R. S. Bradley, and M. K. Hughes (1999), Northern Hemisphere temperatures during the last millennium: Inferences, uncertainties, and limitations, *Geophys. Res. Lett.*, *26*, 759–762.

Morimoto, M., O. Abe, H. Kayanne, N. Kurita, E. Matsumoto, and N. Yoshida (2002), Salinity records for the 1997–98 El Niño from Western Pacific corals, *Geophys. Res. Lett.*, *29*(11), 1540, doi:10.1029/2001GL013521.

Southon, J., M. Kashgarian, M. Fontugne, B. Metiver, and W. W. S. Yim (2002), Marine reservoir corrections for the Indian Ocean and Southeast Asia, *Radiocarbon*, *44*, 167–180.

Stothers, R. B. (1984), The great Tambora eruption in 1815 and its aftermath, *Science*, *224*, 1191–1198.

Stott, L. K., C. Poulsen, S. Lund, and R. Thunell (2002), Super ENSO and global climate oscillations at millennial time scales, *Science*, *297*, 222–226.

Stott, L., K. Cannariato, R. Thunell, R. Haug, A. Koutavas, and S. Lund (2004), Decline of surface temperature and salinity in the western tropical Pacific Ocean in the Holocene epoch, *Nature*, *431*, 56–59.

Stuiver, M., and P. J. Reimer (1993), Extended ^{14}C data base and revised CALIB 3.0 ^{14}C age calibration program, *Radiocarbon*, *35*, 215–230.

Sun, D. Z., J. Fasullo, T. Zhang, and A. Roubicek (2003), On the radiative and dynamical feedbacks over the equatorial Pacific cold tongue, *J. Clim.*, *16*, 2425–2432.

Treydte, K. S., A. H. Schleser, G. Helle, D. C. Frank, M. Winiger, G. H. Haug, and J. Esper (2006), The twentieth century was the wettest period in northern Pakistan over the past millennium, *Nature*, *440*, 1179–1182.

Verschuren, D., K. R. Laird, and B. F. Cumming (2000), Rainfall and drought in equatorial East Africa during the past 1,100 years, *Nature*, *403*, 410–414.

Visser, K., R. Thunell, and L. Stott (2003), Magnitude and timing of temperature change in the Indo-Pacific warm pool during deglaciation, *Nature*, *412*, 152–155.

Wang, L., M. Sarnthein, H. Erlenkeuser, P. M. Grootes, J. O. Grimalt, C. Pelejero, and G. Linck (1999), Holocene variations in Asian Monsoon moisture: A bidecadal sediment record from the South China Sea, *Geophys. Res. Lett*, *26*, 2889–2892.

Wantanabe, T., A. Winter, and T. Oba (2001), Seasonal changes in sea surface temperature and salinity during the Little Ice Age in the Caribbean Sea deduced from Mg/Ca and $^{18}O/^{16}O$ ratios, *Mar. Geol.*, *173*, 21–35.

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