Millennial meridional dynamics of Indo-Pacific Warm Pool during the last termination

Li Lo¹, Chuan-Chou Shen¹*, Kuo-Yen Wei¹, George S. Burr¹,², Horng-Sheng Mii³, Min-Te Chen⁴, Shih-Yu Lee⁵, Meng-Chieh Tsai¹

¹High-Precision Mass Spectrometry and Environment Change Laboratory (HISPEC), Department of Geosciences, National Taiwan University, Taipei 10617, Taiwan ROC
²NSF-Arizona Accelerator Mass Spectrometry Facility, Department of Physics, University of Arizona, Tucson, AZ 85721, USA
³Department of Earth Sciences, National Taiwan Normal University, Taipei 11677, Taiwan ROC
⁴Institute of Applied Geosciences, National Taiwan Ocean University, Keelung 20224, Taiwan ROC
⁵Research Center for Environmental Changes, Academia Sinica, Taipei 11529, Taiwan ROC

Revised to Climate of the Past

2014.11.23

*Corresponding Author: Chuan-Chou Shen

Email: river@ntu.edu.tw; Tel: 886-2-3366-5878; Fax: 886-2-3365-1917
Abstract

To develop an in-depth understanding of the natural dynamics of the Indo-Pacific Warm Pool (IPWP) during the last deglaciation, stacked North- (N-) and South-IPWP (S-IPWP) thermal and hydrological records over the past 23-10.5 thousand years (ka) were built using planktic foraminiferal geochemistry data from a new core, MD05-2925 (9.3°S, 151.5°E, water depth 1661 m) in the Solomon Sea and eleven previous sites. Ice-volume corrected seawater δ¹⁸O (δ¹⁸O_SW-IVC) stacks show that S-IPWP δ¹⁸O_SW-IVC values are indistinguishable from their northern counterpart through glacial time. The N-IPWP SST stacked record features an increasing trend of 0.5 °C ka⁻¹ since 18 ka. Its S-IPWP counterpart shows an earlier onset of temperature increase at 19 ka and a strong teleconnection to high-latitude climate in the Southern Hemisphere. Meridional SST gradients between N- and S-IPWP were 1 to 1.5 °C during the Bølling/Allerød period and < 0.5 °C during both Heinrich event 1 and the Younger Dryas due to a warmer S-IPWP. A warm S-IPWP during the cold events may possibly weaken the southern hemispheric branch of the Hadley Cell and reduce precipitation in the Asian Monsoon region.
1. Introduction

The Indo-Pacific Warm Pool (IPWP) is the largest warm water mass in the world, with an annual average sea surface temperature (SST) greater than 28 °C (Yan et al., 1992). Vigorous regional atmosphere circulation transports latent heat and water moisture from the IPWP to the middle and high latitudes (Yan et al., 1992). For the past five decades, the IPWP has experienced surface water freshening and a westward shift in precipitation, resulting in regional drought in East Africa and storm track changes in East Australia (Cravatte et al., 2009; Williams and Funk, 2011). Since the early 2000s, intensive paleoclimatological studies have been conducted to understand long-term thermal and hydrological changes in the IPWP, associated with glacial/interglacial (G/IG) cycles, and to constrain the relationship between warm pool thermal and hydrological fluctuations to high latitude ice sheet and greenhouse gas concentrations during the late Pleistocene (e.g., Lea et al., 2000; Stott et al., 2002; Visser et al., 2003; Rosenthal et al., 2003; Stott et al., 2004; de Garidel-Thoron et al., 2005; Steinke et al., 2006; Levi et al., 2007; Xu et al., 2008; Linsley et al., 2010; Bolliet et al., 2011; Mohtadi et al., 2014).

Stacked IPWP SST and seawater oxygen isotope ($\delta^{18}O_{SW}$) records from the last glacial to the Holocene clearly show a close link between the IPWP SST, the Asian-Australian Monsoon (AAM) system, and sea level (Stott et al., 2004; Oppo et al., 2009; Linsley et al., 2010). However, a complicated ocean-island configuration and regional topography hinder the fidelity of using these records to describe past climate changes in detail (Griffiths et al., 2009; Mohtadi et al., 2011). In particular, little is known about the meridional
thermal-hydrological dynamics between the N-IPWP and S-IPWP during the last termination.

Here we present new oceanic proxy-inferred SST and ice volume-corrected surface seawater oxygen isotope $\delta^{18}$O ($\delta^{18}$O$_{SW-IVC}$) records from the Solomon Sea, Papua New Guinea (PNG) for the past 23-10.5 thousand years ago (ka, before 1950 AD, hereafter). New SST and $\delta^{18}$O$_{SW-IVC}$ stacked records since the last termination are built for both the N- and S-IPWP to understand regional thermal-hydrological changes and interhemispheric teleconnections.

2. Material and Methods

Site MD05-2925 (9.3°S, 151.5°E, water depth 1661 m) is located at the northern slope of the Woodlark Basin in the Solomon Sea, which is the passage of surface and subsurface water masses between low- and middle-latitude South Pacific Ocean gyre and cross equatorial currents (Grenier et al., 2011; Melet et al., 2011) (Fig. 1). The seasonal precipitation in this region (Fig. 1) is dominated by the AAM system, coupled with the intertropical convergence zone (ITCZ) (Shiau et al., 2012, and references therein). Tests of single species planktonic foraminifera, *Globigerinoides sacculifer* (> 500 µm, total amount of 2-6 mg), at 13 selected depths were picked for accelerator mass spectrometry (AMS) $^{14}$C dating. The AMS dates were calibrated using the CALIB 6.0.1 program (Stuiver et al., 2010, Table 1; Reimer et al., 2009) to reconstruct an age model for a time interval from 23 to 10.5 ka.

Forty to sixty individuals of the planktonic foraminifera *Globigerinoides ruber* (white, s.s., 250-300 µm) were picked under the microscope. For Mg/Ca measurements, 20-30 individuals were gently crushed and transported into a
1.5 mL Teflon vial. The cleaning procedure was as follows: (1) foraminiferal fragments were immersed in ethanol, (2) a 0.45 mL aliquot of 3% H$_2$O$_2$, (3) NH$_4$Cl (0.45 mL, 1.0 N), (4) NH$_2$OH (0.45 mL, 0.01 N), and then (5) dilute nitric acid (1 mL, 0.005 N). A sector field inductive coupled plasma mass spectrometer (SF-ICP-MS), Thermo Electron Element II, housed at the High-Precision Spectrometry and Environment Change Laboratory (HISPEC), Department of Geosciences, National Taiwan University, was used to determine trace element/Ca ratios following the methodology developed by Shen et al. (2007). The detailed cleaning procedure and methodology are available in Lo et al. (2014). Two-year 1-sigma reproducibility of Mg/Ca analyses is ±0.21% (Lo et al., 2014). We used a composite Mg/Ca-SST equation by Anand et al. (2003) to calculate SSTs.

For oxygen stable isotope analysis, 7-10 individuals were immersed in methanol, ultrasonicated for 10 seconds, and then rinsed with deionized water 5 times. Samples were immersed afterward in a hyperchloride sodium (NaOCl) for 24 hours, and then analyzed with an isotopic ratio mass spectrometer (IRMS), Micromass IsoPrime, housed in the National Taiwan Normal University. Long-term 1-sigma precision is better than ±0.05‰ (N = 701, Lo et al., 2013) with respect to Vienna Pee Dee Belemnite (VPDB).

To extract seawater δ$^{18}$O (δ$^{18}$O$_{SW}$) values, we used a cultural based equation, SST = 16.5 - 4.8 × (δ$^{18}$O$_C$ - δ$^{18}$O$_{SW}$) (Bemis et al., 1998) and a constant offset of 0.27‰ between carbonate VPDB and Vienna Standard Ocean Water (VSMOW) scales. Ice volume corrected δ$^{18}$O$_{SW}$ (δ$^{18}$O$_{SW-IVC}$) was calculated using the method proposed by Waelbroeck et al. (2002).
The empirical orthogonal function (EOF) analysis of a modern SST dataset (1950-2004 AD, Reynolds et al., 2002) for a sector from 20°S – 20°N, and 100°E- 180°E was conducted (Fig. 2) to determine the boundary between N- and S-IPWP. With an equatorial border, the EOF1 factor (83.4%) clearly resolved different SST variation groups. The EOF2 factor shows minor (9.7%) but significant inter-annual zonal (ENSO) control on the SST patterns. EOF results show that the geographic equator is also the thermal equator between N-IPWP and S-IPWP (Fig. 2).

To build a stacked N- and S-IPWP record, we followed the suggestions by Leduc et al. (2010) and considered three criteria for this dataset: (1) sites with locations from 12°N to 15°S, which is the main IPWP range (Yan et al., 1992; Gagan et al., 2004), and (2) usage of specific proxies, Mg/Ca-derived SST and δ18O records of planktonic foraminifera, G. ruber (white, s.s.). Records from 12 sites were selected, including this study (Table 2). We adopted the published age model for sites ODP806, MD97-2140, MD97-2141, MD98-2162, MD98-2170, MD98-2176, and MD98-2181. For records with available original radiocarbon ages from sites, including MD01-2378, MD01-2390, MD98-2165, and MD06-3067, we recalculated the age models using the CALIB 6.0.1 program. The sea level change effect on δ18Osw was also corrected. We divided the total data into 400-yr windows and calculated the mean and standard error of the mean for each time window.

3. Results and Discussion

3.1 Geochemical proxy data at site MD05-2925
Planktonic foraminiferal geochemical proxy data for site MD05-2925 are shown in Figure 3. *G. ruber* δ¹⁸O varies from -1.0 to -2.3‰ and shows no significant millennial timescale variations. Mg/Ca ratios feature stable glacial values of ~3.5 mmol/mol and rapid increasing transitions of 0.5-1.0 mmol/mol at ~18.5, 16.5, 14.5, and 12.8 ka. The glacial-interglacial variation of calculated seawater δ¹⁸O changes is ~1‰. Two abrupt decreases of 0.6-0.8‰ are observed at 14.6 and 11.8 ka.

### 3.2 Solomon SST and δ¹⁸O<sub>SW-IVC</sub> records during the last termination

Mg/Ca SST records of the planktonic foraminifera *G. ruber* reveal a stable glacial thermal condition during the period 23.0-18.5 ka, with a variation <1 °C and a glacial-interglacial difference of ~3 °C between the last glacial maximum (LGM) and the end of the Younger Dryas (YD) in the Solomon Sea (Fig. 4A). This record is characterized by (i) the end of glacial conditions at 18.5 ka, and (ii) rapid SST increases of 1-2 °C at 18.5-18.0, 17.0-16.0, 15.0-14.5, and 13.0-12.5 ka.

The onset of deglacial SST increases in this region is consistent with the timing of thermal changes in the Southern Ocean as inferred from Antarctic ice core δD records (Stenni et al., 2003) (Fig. 4A). This agreement indicates a strong climatic teleconnection between low- and high-latitude realms in the Southern Hemisphere (SH), as well as change of greenhouse gas concentrations (Mothadi et al., 2014). There are significant SST increases of 1-2 °C during Heinrich event (H1) and the YD. Previous studies from the Eastern Equatorial and South Pacific reveal a mechanism characterized by early warming of South Pacific subtropical mode water (Pahnke et al., 2003;...
This warm signal is transported along a gyre to the east equatorial Pacific (EEP) and eventually to the west Pacific through ocean tunneling (Pena et al., 2008; Qu et al., 2013, Fig. 4A). Our new SST record is similar to those in the EEP (Pena et al., 2008) and eastern Indian Ocean records (Xu et al., 2008; Mothadi et al., 2014) for both termination timing (within dating error) and significant warming during the H1 and YD events. There is a slightly warming (<1 °C) interval at 14.5-13.5 ka during the B/A period (Fig. 4A). The warming could be attributed to a possible mixing with the warm N-IPWP surface water.

The Solomon Sea δ18O_{SW-IVC} record is given in Figure 4B. It varies from -0.5 to 0.1‰ during 23.0-10.5 ka. A relatively stable condition with 1-sigma variability of 0.1‰ occurred from 23.0 to 16.0 ka. Two significant positive excursions with 0.2-0.5‰ enrichments in δ18O are observed in the intervals 16.8-15.0, and 13.8-11.8 ka. Two stable periods with low δ18O_{SW-IVC} of -0.4‰ occurred between 15.0-13.0 ka and after 11.8 ka.

The dramatic δ18O_{SW-IVC} increases during H1 and the YD likely resulted from a weakening and/or southward shift of the ITCZ (Chiang and Bitz, 2005; Broccoli et al., 2006), and local evaporation may also play a role. Agreement of δ18O sequences of Greenland NGRIP ice core and the Solomon Sea δ18O_{SW-IVC} indicates an imprint from high latitude Northern Hemisphere (NH) during the last termination period (Shakun and Carlson, 2010) (Fig. 4B).

### 3.3 Millennial timescale variations of N- and S-IPWP SST stacks

Both N- and S-IPWP stacked SSTs show the same difference of ~3 °C between the last glacial and interglacial states (Fig. 5A). N-IPWP stacked SST
values increased steadily since 18 ka through the termination at a rate of 0.5 
°C/kyr. Millennial timescale variability is absent in this record, which is similar
to Linsley et al. (2010) and Stott et al. (2002). Although the resolution of ODP
806 and MD97-2140 are less than our request to solve millennial-timescale
event, there is no significant difference with/without their records in our N-
IPWP stacks (not shown).

The onset of the termination at ~19 ka in the S-IPWP stack is consistent
with temperature increases in Antarctica (Stenni et al., 2003), and occur about
1 kyr earlier than in the N-IPWP stack (Fig. 5A). This timing is synchronous
with EEP (Pena et al., 2008) and non-upwelling region eastern Indian Ocean
(Xu et al., 2008; Mothadi et al., 2014) SST records. Thus, our MD05-2925 and
S-IPWP stacked SST may not severely controlled by the equatorial upwelling
intensity. Instead of that, S-IPWP stacked SST represents broad SH
equatorial region thermal conditions under upwelling/non-upwelling, E-W
equatorial and even in the different ocean basin (Indian/Pacific Ocean). The
S-IPWP stacked SST record is characterized by a warming trend during H1
and the YD periods, similar to Antarctic ice core temperature records (Stenni
et al., 2003), and a steady thermal condition at ~27 °C during Bølling/Allerød
(B/A), corresponding to the Antarctic Cold Reversal (ACR) (Fig. 5A).

The thermal gradient between N- and S-IPWP is around 1 °C during 23 to
19 ka. Due to the earlier S-IPWP warming, the thermal gradient dropped from
1 to 0.5 °C around 19-18 ka, and persisted to the end of the H1 event. The
largest observed thermal gradient (1.5-2.0 °C) occurred during the B/A period,
and was followed by a 1 °C drop during the YD. The meridional SST gradient
between N- and S-IPWP over the last termination is attributed to the large
thermal variability in the S-IPWP (Fig. 5A). Asynchroneity between persistent
N-IPWP and fluctuating S-IPWP SST sequences (Fig. 5A) indicates a
meridionally dynamic IPWP through the last termination period. This N-S SST
gradient variability would also affect interhemispheric air flow and heat
transport (Gibbons et al., 2014; McGee et al., 2014), providing a mechanism
to explain heat transport between the hemispheres on a millennial timescale.

3.4 N- and S-IPWP $\delta^{18}$O_{SW-IVC} records

Both N- and S-IPWP $\delta^{18}$O_{SW-IVC} records feature (i) low values of -0.3-0.0‰
during glacial times, and (ii) increasing trends after 19 ka (Fig. 5C). The
gradient between N- and S-IPWP gradually increased from 0‰ to 0.2‰
through the termination (Fig. 5D). A similar pattern of $\delta^{18}$O_{SW-IVC} between N-
and S-IPWP suggests that hydrological conditions in the two regions were
governed by the same factor(s), probably related to Northern Atlantic cold
perturbations (Shakun and Carlson, 2010). It has also been suggested that a
major $\delta^{18}$O_{SW-IVC} increase during the H1 and YD periods in the IPWP region
likely resulted from reduced precipitation and oceanic advection in both the N-
IPWP and S-IPWP regions (Gibbons et al., 2014; McGee et al., 2014).

3.5 Meridional IPWP SST gradient and the southward-shifted ITCZ
precipitation boundary

A striking feature of the stacked SST records is the warming in the S-IPWP
during the H1 and YD periods (Fig. 5A). Observations over the past six
decades (Fig. 12 of Feng et al., 2013) show that an equatorward shift of the
NH convection branch of the Hadley Cell (HC) could result from an oceanic
warming at ~10° S. This equatorward shift could induce a southward ITCZ shift of about 10° (Feng et al., 2013). Model simulations (Chiang and Bitz, 2005; Broccoli et al., 2006, Lee et al., 2011) suggest that this altered circulation is a powerful teleconnection between the NH and SH climate systems through a coupled tropical ocean-atmosphere pathway, and is supported by marine and terrestrial hydrological proxy data (Wang et al., 2001, Lea et al., 2003, Wang et al., 2007, Griffiths et al., 2009, Shakun and Carlson, 2010, Mohtadi et al., 2011, Meckler et al., 2012, Ayliffe et al., 2013, Carolin et al., 2013, Gibbons et al., 2014; McGee et al., 2014, Fig. 6).

Distinctly different precipitation conditions across 8-10°S in the IPWP during the H1 and YD events are illustrated in Figure 6. For example, enhanced terrestrial sediment flux into the Coral Sea is suggested by a marine sediment thorium isotopic proxy record at 11° S (Shiau et al. 2011). Lynch’s crater records from northeastern Australia at 17° S (Muller et al., 2008) show strong Australian summer monsoonal conditions. Stalagmite δ18O records at Flores Island (8° S) also feature intense precipitation during H1 and the YD (Griffiths et al., 2009, Ayliffe et al., 2013). However, marine and stalagmite δ18O evidence reveal conditions of reduced precipitation and increased salinity in the northern IPWP north of 8-10° S, including the South China Sea (12° N, Stenike et al., 2006), Sulu Sea (8° N, Rosenthal et al., 2003), Philippine Sea (6° N, Stott et al., 2002; Boillet et al., 2011), Java Island (8° S, Mohtadi et al., 2011), Solomon Sea (9° S, this study), and Borneo island (4° N, Meckler et al., 2012, Carolin et al., 2013) (Fig. 6). On the basis of previous terrestrial and marine hydrological records and our new data, as well as modern (Feng et al., 2013) and simulated (Chiang and Bitz, 2005: Broccoli
et al., 2006) data, we speculate a sharp precipitation boundary between the maritime continents and Australia at about 8-10° S, extending from the Solomon Sea, Arafura Sea and Timor Sea, to the eastern Indian Ocean during H1 and the YD periods (Fig. 6). We propose that the west and east boundaries are between the Java-Flores islands (Griffiths et al., 2009, Mohtadi et al., 2011), and Solomon-Coral Seas, respectively (Shiau et al., 2011, this study). Geographical pattern mismatch between thermal and precipitation could be associated with the local convection branch shifting and sea level change (Linsley et al., 2010).

To sum up our geochemical and composite dataset in the IPWP region during the last terminations, we propose that the enlarged IPWP meridional SST gradient could result in an altered HC and reduced (increased) precipitation for the East Asian (Australia) monsoon territories during the H1 and YD periods (McGee et al., 2014). We also propose that variations in the meridional IPWP SST gradient during the termination period were mainly caused by the S-IPWP, which is closely linked to high-latitude climate systems.

4. Conclusions

Our new MD05-2925 marine geochemical records and previous reports suggest that the meridional IPWP thermal conditions are strongly linked to interhemispheric high-latitude climate during the last deglaciation. Ice volume-corrected δ¹⁸O_sw stacked records show an increasing salinity gradient between N- and S-IPWP over the last termination. Here we propose a new process of the thermal evolution of IPWP region, which meridional differences
of its thermal gradient could amplify the signal from high latitude Northern
hemisphere climate events and radiative forcing from greenhouse gases. A
hypothetical precipitation boundary around 8-10° S during H1 and the YD has
also been proposed, which is most likely caused by the meridional IPWP SST
gradient and HC anomalies. More advanced high-resolution regional model
simulations are required to clarify (1) local precipitation variation in response
to the complicated sea level and convection change, (2) the role of IPWP
meridional thermal-hydrological gradient to an altered HC, and (3) its
relationship with regional and global climate systems during global climate
perturbation events.
Acknowledgements

MD05-2925 site location was selected by Min-Te Chen and Meng-Yang Lee and collected during the IMAGES PECTEN Cruise, conducted by Luc Beaufort and Min-Te Chen. Chien-Ju Chou, Wan-Lin Hu, and Yu-Ting Hsiao helped to pick foraminifera samples. Yang-Hui Hsu helped to operate the climatological database and plotted figures. Thanks to Delia W. Oppo and Braddock K. Linsley for their generous offering of the non-overlapping method MatLab code. This research was funded by Taiwan ROC MOST (99-2611-M-002-005, 100-2116-M-002-009 and 103-2119-M-002-022 to CCS; 95-2611-M-002-019 and 96-2611-M-002-019 to KYW), and National Taiwan University (101R7625 to CCS).
References


Griffiths, M. L., Drysdale, R. N., Gagan, M. K., Zhao, J.-X., Ayliffe, L. K.,
Hellstrom, J. C., Hantoro, W. S., Frisia, S., Feng, Y.-X., Cartwright, I., St.
Pierre, E., Fischer, M., J., and Suwargadi, B. W.: Increasing Australian-
Indonesian monsoon rainfall linked to early Holocene sea-level rise. Nature

Antarctic timing of surface water changes off Chile and Patagonian ice

Lea, D. W., Pak, D. K., and Spero, H. J.: Climate impact of late Quaternary
equatorial Pacific sea surface temperature variations. Science, 289, 1719-

Lea, D. W., Pak, D. K., Peterson, L. C., and Hughen, K. A.: Synchronicity of
tropical high-latitude Atlantic temperatures over the last glacial termination.

Leduc, G., Schneider, R., Kim, J.H., and Lohmann, G.: Holocene and Eemian
sea surface temperature trends as revealed by alkenone and Mg/Ca

Levi, C., Labeyrie, L., Bassinot, F., Guichard, F., Cortijo, E., Waelbroeck, C.,
Caillon, N., Duprat, J., de Garidel-Thoron, T., and Elderfield, H.: Low-
latitude hydrological cycle and rapid climate changes during the last
deglaciation. Geochem., Geophys., Geosy., 8(5), Q05N12, doi:

Linsley, B. K., Rosenthal, Y., and Oppo, D. W.: Holocene evolution of the
Indonesian throughflow and the western Pacific warm Pool. Nature Geosci.,

Lo, L., Lai, Y.-H., Wei, K.-Y., Lin, Y.-S., Mii, H.-S., Shen, C.-C.: Persistent sea
surface temperature and declined sea surface salinity in the northwestern
tropical Pacific over the past 7500 years. J. Asian Earth Sci., 66, 234-239,
2013.

Lo, L., Shen, C.-C., Lu, C.-J., Chen, Y.-C., Chang, C.-C., Wei, K.-Y., Qu, D.,
and Gagan, M. K.: Determination of element/Ca ratios in foraminifera and
corals using cold- and hot-plasma techniques in inductively coupled plasma

location and cross-equatorial heat transport at the Last Glacial Maximum,
Heinrich Stadial 1, and the mid-Holocene. Earth Planet. Sci. Lett., 390, 69-
79, 2014.

Meckler, A. N., Clarkson, M. O., Cobb, K. M., Sodemann, H., and Adkins, J. F.:
Interglacial hydroclimate in the tropical West Pacific through the late


Shakun, J. D., Clark, P. U., He, F., Marcott, S. A., Mix, A. C., Liu, Z., Otto-Bliesner, B., Schmittner, A., and Bard, E.: Global warming preceded by increasing carbon dioxide concentrations during the last deglaciation. Nature, 484, 49-55, 2012.


Table 1 AMS $^{14}$C dates of site MD05-2925.

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>$^{14}$C ages (years)</th>
<th>Error (years)</th>
<th>Cal. ages (years)</th>
<th>Error (years)</th>
</tr>
</thead>
<tbody>
<tr>
<td>117</td>
<td>8823</td>
<td>50</td>
<td>9414</td>
<td>111</td>
</tr>
<tr>
<td>127*</td>
<td>10306</td>
<td>70</td>
<td>11259</td>
<td>159</td>
</tr>
<tr>
<td>140</td>
<td>10441</td>
<td>30</td>
<td>11333</td>
<td>80</td>
</tr>
<tr>
<td>147*</td>
<td>11477</td>
<td>70</td>
<td>12854</td>
<td>110</td>
</tr>
<tr>
<td>157</td>
<td>12066</td>
<td>60</td>
<td>13391</td>
<td>84</td>
</tr>
<tr>
<td>172*</td>
<td>13117</td>
<td>70</td>
<td>14973</td>
<td>309</td>
</tr>
<tr>
<td>180</td>
<td>13748</td>
<td>35</td>
<td>16283</td>
<td>453</td>
</tr>
<tr>
<td>192*</td>
<td>14080</td>
<td>74</td>
<td>16746</td>
<td>223</td>
</tr>
<tr>
<td>207*</td>
<td>15616</td>
<td>75</td>
<td>18201</td>
<td>175</td>
</tr>
<tr>
<td>217</td>
<td>16470</td>
<td>81</td>
<td>19083</td>
<td>90</td>
</tr>
<tr>
<td>262*</td>
<td>18985</td>
<td>94</td>
<td>22167</td>
<td>181</td>
</tr>
<tr>
<td>272*</td>
<td>20960</td>
<td>150</td>
<td>24411</td>
<td>167</td>
</tr>
<tr>
<td>292*</td>
<td>21650</td>
<td>78</td>
<td>25304</td>
<td>339</td>
</tr>
</tbody>
</table>

*Samples were measured in the NSF-Arizona AMS Laboratory of the University of Arizona (U. Arizona), Tucson, USA, and the others were measured in the Rafter Radiocarbon Laboratory, Institute of Geological and Nuclear Science (GNS), New Zealand.
Table 2 Selected sites for stacked N- and S-IPWP records.

<table>
<thead>
<tr>
<th>Core</th>
<th>Location (Latitude, and longitude)</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>North-IPWP group</strong> (orange circles in Figs 1 and 2)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ODP 806</td>
<td>0.3°N, 159.4°E</td>
<td>Lea et al. (2000)</td>
</tr>
<tr>
<td>MD97-2140</td>
<td>2.0°N, 141.7°E</td>
<td>de Garidel-Thoron et al. (2005)</td>
</tr>
<tr>
<td>MD98-2181</td>
<td>6.3°N, 125.8°E</td>
<td>Stott et al. (2002, 2004)</td>
</tr>
<tr>
<td>MD06-3067</td>
<td>6.5°N, 126.5°E</td>
<td>Bolliet et al. (2011)</td>
</tr>
<tr>
<td>MD97-2141</td>
<td>8.8°N, 121.3°E</td>
<td>Rosenthal et al. (2003)</td>
</tr>
<tr>
<td>MD01-2390</td>
<td>12.1°N, 113.2°E</td>
<td>Stenike et al. (2006)</td>
</tr>
<tr>
<td><strong>South-IPWP group</strong> (green circles and star in Figs 1 and 2)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>MD98-2162</td>
<td>4.4°S, 117.5°E</td>
<td>Visser et al. (2003)</td>
</tr>
<tr>
<td>MD98-2176</td>
<td>5.0°S, 133.4°E</td>
<td>Stott et al. (2004)</td>
</tr>
<tr>
<td>MD05-2925</td>
<td>9.3°S, 151.5°E</td>
<td>This Study</td>
</tr>
<tr>
<td>MD98-2165</td>
<td>9.7°S, 118.3°E</td>
<td>Levi et al. (2007)</td>
</tr>
<tr>
<td>MD98-2170</td>
<td>10.6°S, 125.4°E</td>
<td>Stott et al. (2004)</td>
</tr>
<tr>
<td>MD01-2378</td>
<td>13.1°S, 121.7°E</td>
<td>Xu et al. (2008)</td>
</tr>
</tbody>
</table>
Figure captions

Fig. 1. Climatological map of the Indo-Pacific Warm Pool (IPWP) sea surface temperature (SST, left) and precipitation (right) during 1950-2004 AD (Reynolds et al., 2002). Upper panels are June-July-August (JJA), and lower panels are December-January-February (DJF) averages of (A, C) SSTs and (B, D) precipitation distribution maps. SST and precipitation are at 0.5 °C and 2 mm/day intervals. Our study site MD05-2925 is shown as the green star. Orange and green dots denote previous study sites in the IPWP region (Table 2) for reconstruction of meridional thermal and precipitation variations during the glacial/interglacial change.

Fig. 2. EOF analysis on SST (Dataset from Reynolds et al., 2002) and selected sites (Table 2) used for stacked N- and S-IPWP records. (A) EOF1 explains 83.4% of the total variance, which mainly represents intra-annual seasonality. (B) EOF2 shows a clear zonal pattern. Orange circles represent selected sites for the N-IPWP group and green ones for the S-IPWP group. The green star denotes the MD05-2925 site used in this study.

Fig. 3. Planktonic foraminifera G. ruber geochemical proxy records of site MD05-2925, including (A) oxygen isotope (δ¹⁸O), (B) Mg/Ca ratio, and (C) temperature corrected-only seawater oxygen isotope (δ¹⁸O_SW). Triangle symbols are corrected radiocarbon dates (Table 1).

Fig. 4. Geochemical proxy records of MD05-2925. (A) SST (red circles and line) and (B) δ¹⁸O_SW-IVC (blue line) were reconstructed with G. ruber Mg/Ca ratios and δ¹⁸O. The cyan line denotes the Antarctica EPICA deuterium isotope record (Stenni et al., 2003), and the yellow line is the Greenland ice core NGRIP (Northern Greenland Ice Core Project Members, 2004) oxygen isotope record. The superimposed dark cyan and dark yellow lines are the 200-yr smoothed records, respectively. Black triangles are AMS ¹⁴C dates (Table 1). Vertical bars denote the H1 and YD periods.

Fig. 5. Four hundred-year non-overlapping binned (A) SST and (C) δ¹⁸O_SW-IVC of N- (orange solid line) and S-IPWP (green solid line). Lower panel show the differences in (B) SST and (D) δ¹⁸O_SW-IVC between N- and S-IPWP. The compilations of N- and S-IPWP surface water thermal and hydrological records (Table 2) were calculated with the non-overlapping binned methods (Oppo et al., 2009; Linsley et al., 2010). All dashed lines represent 1-sigma uncertainty ranges. Gray bars show the H1 and YD periods.

Fig. 6. Hypothetical proxy-inferred precipitation boundary during the H1 and YD events (modified from the Linsley et al., 2010). Blue dots represent relatively increasing precipitation/δ¹⁸O_SW lighter condition, and brown ones a decreasing precipitation/δ¹⁸O_SW heavier condition. The segment between Java and Flores Islands of this sharp boundary (red dashed line) was proposed by Mohtadi et al. (2011), and the one between the Solomon and Coral Seas by this study. Black contours represent SST.
Fig. 1

Sea surface temperature

Precipitation

---

MD05-2925

---

mm/day
Fig. 4
Seas into the Makassar Strait during the boreal winter monsoon. Salinity data from ref. The grey arrows show the generalized flow of the ITF. The dark green arrows show the flow of relatively fresh water from the South China and Java. Cores discussed in this study are indicated.

SST data from ref. Figure 1

Table 1

<table>
<thead>
<tr>
<th>Core ID</th>
<th>Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>MD90</td>
<td>South China Sea</td>
</tr>
<tr>
<td>MD41</td>
<td>Makassar Strait</td>
</tr>
<tr>
<td>MD67 (?)</td>
<td>Flores Island</td>
</tr>
<tr>
<td>MD81</td>
<td>Java Island</td>
</tr>
<tr>
<td>MD62</td>
<td>Makassar Strait</td>
</tr>
<tr>
<td>MD65</td>
<td>Flores Island</td>
</tr>
<tr>
<td>MD78 (?)</td>
<td>Lynch's Crater</td>
</tr>
<tr>
<td>GeoB10053-7</td>
<td></td>
</tr>
<tr>
<td>MD25</td>
<td>ODP806 (?)</td>
</tr>
</tbody>
</table>

Sediment cores used in this study.

4.4 84

<table>
<thead>
<tr>
<th>Latitude</th>
<th>Longitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>20° N</td>
<td>180° E</td>
</tr>
<tr>
<td>10° N</td>
<td>180° E</td>
</tr>
<tr>
<td>0°</td>
<td>180° E</td>
</tr>
<tr>
<td>20° S</td>
<td>180° E</td>
</tr>
<tr>
<td>10° S</td>
<td>180° E</td>
</tr>
<tr>
<td>0°</td>
<td>180° E</td>
</tr>
</tbody>
</table>

SST a calculated relative to average of SST from nemii–netii yr.

Only Not used in SST a composite reconstructions aMDti and kiGGC bg

MDikflmtic

MDtslkfrpe Banda Sea

MDtrlkfnke Sulu Sea

MDiklmfrse Timor Sea

MDtslkfpoe Sumbae Indonesia MDpo

MDtslkfple Sg Makassar Stg MDpl n

riGGCe SW Sulawesie Makassar Stg riGGC m

kiGGCc

Core IDs location

Fig. 6