Title

Glacial-interglacial changes in equatorial Pacific surface-water structure during the Plio-Pleistocene intensification of Northern Hemisphere Glaciation

Authors

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Abstract

The late Pliocene/early Pleistocene intensification of Northern Hemisphere Glaciation (iNHG) represents the most recent major tipping point in Earth's climate history. In the equatorial Pacific, the iNHG has been suggested to mark the transition from a warm Pliocene climate state, in which the Pacific warm pool was greatly expanded relative to today giving rise to a surface ocean temperature field akin to modern El-Niño conditions, to a climate state with strong upwelling in the east of the basin and stronger zonal temperature gradients. Here we examine this transition by presenting new high-resolution (~800 yr) planktic foraminiferal Mg/Ca- and $\delta^{18}\text{O}$ -based sea-surface temperature (SST) and salinity (SSS; based on $\delta^{18}\text{O}$ residuals [$\delta^{18}\text{O}_{\text{ivc-sw}}$]) records from Eastern Equatorial Pacific (EEP) Ocean Drilling Program Site 849 spanning ~2.75–2.4 Ma. Our study interval contains the first large-amplitude (~1‰ in benthic $\delta^{18}\text{O}$) early Pleistocene glacial-interglacial cycles (Marine Isotope Stages 100–96) representing the culmination of iNHG. In this study we investigate west-to-east equatorial Pacific SST and SSS gradients. We find that the zonal SST gradient increased by ~1.5°C from ~2.75 to ~2.4 Ma through EEP cooling. No glacial-interglacial pattern is seen in this gradient. On the other hand, we document obliquity-paced oscillations in the $\delta^{18}\text{O}_{\text{ivc-sw}}$ (SSS) gradient. We hypothesize that these oscillations originated from a response to a glacial southward and interglacial northward migration of the Intertropical Convergence Zone over the EEP.

1. Introduction

A number of influential studies of the long-term evolution of the tropical Pacific mean state have suggested that the warm Pliocene equatorial Pacific Ocean was characterized by a “perennial El-Niño-like state” (e.g., Wara et al., 2005; Fedorov et al., 2006) and that the intensification of Northern Hemisphere Glaciation (iNHG) was associated with a gradual increase in west-to-east (zonal) equatorial Pacific sea-surface temperature (SST) gradients (Wara et al., 2005). However, this interpretation is a matter of ongoing debate (Ravelo et al., 2014; Zhang et al., 2014). Also debated is the suggestion that fluctuations took place between El Niño- and La Niña-like climate states on glacial-interglacial (G-IG) timescales across iNHG (e.g., Philander and Fedorov, 2003; Fedorov et al., 2006; Bolton et al., 2010).

The iNHG, typically considered to span the interval ~3.6 to ~2.4 Ma (Mudelsee and Raymo, 2005), is part of a strong long-term cooling trend following the Mid-Piacenzian warm period. Of particular significance during iNHG is Marine Isotope Stage (MIS) 100 (~2.5 Ma), which is suggested to be the earliest glacial during which the Laurentide Ice Sheet advanced into the mid-latitudes (Balco and Rovey, 2010). Advance and retreat of the Laurentide Ice Sheet had important consequences for ice rafting in the North Atlantic (Bailey et al., 2013). It is also hypothesized to have influenced trade-wind strength and thus the strength of upwelling in the Eastern Equatorial Pacific (EEP) during glacial terminations (Jakob et al., 2016). These observations highlight the need to improve our understanding of obliquity-controlled surface-water dynamics including Walker Cell circulation in the tropical Pacific during iNHG. The Pacific Walker Cell circulation comprises easterly and westerly winds at the lower and upper troposphere, and rising motion and subsidence over the western and eastern tropical Pacific, respectively (McPhaden et al., 2006). This circulation is weaker during El Niño states and stronger during La Niña states of the El Niño-Southern Oscillation (ENSO) dynamics (McPhaden et al., 2006). Some studies invoke strong changes in ocean-atmosphere dynamics related to Walker Cell circulation (Philander and Fedorov, 2003; Fedorov et al., 2006; Zhang

et al., 2014) to account for G-IG variability in zonal SST gradients in the Plio-Pleistocene Pacific Ocean. Other studies invoke ocean-atmosphere teleconnections to southern high latitudes rather than local fluctuations between El Niño- and La Niña-like climate states as a driver for tropical G-IG climate fluctuations (de Garidel-Thoron et al., 2005; Lee and Poulsen, 2005; Lawrence et al., 2006; Bolton et al., 2010; Jakob et al., 2016).

To further investigate whether changes in Walker Cell circulation may have driven large-scale changes in tropical Pacific Ocean conditions, particularly west-east thermocline-tilt-related changes in surface-water characteristics, during the late Pliocene to early Pleistocene iNHG, we present high-resolution planktic foraminiferal Mg/Ca and $\delta^{18}\text{O}$ records from EEP Ocean Drilling Program (ODP) Site 849 (Figure 1) covering the time interval ~2.75–2.4 Ma (MIS G6–95). These new records complement benthic and planktic foraminiferal $\delta^{13}\text{C}$ records and a productivity record for MIS G1–95 from the same site presented in Jakob et al. (2016). In the current study, we compare our new Mg/Ca-temperature and $\delta^{18}\text{O}$ records with previously published records from ODP Site 846 in the EEP upwelling tongue (Lawrence et al., 2006), ODP Site 1241 in the East Pacific Warm Pool (Groeneveld et al., 2014), and ODP Site 806 in the West Pacific Warm Pool (WPWP) (Medina-Elizalde and Lea, 2010). We calculate zonal equatorial Pacific SST (ΔSST) and ice volume-adjusted $\delta^{18}\text{O}$ seawater ($\Delta\delta^{18}\text{O}_{\text{ivc-sw}}$) gradients between Western Equatorial Pacific (WEP) Site 806 (data from Medina-Elizalde and Lea, 2010) and EEP Site 849 (this study). These records are all based on the same substrate (planktic foraminifer *Globigerinoides ruber*), thus allowing us to reliably test for obliquity-controlled surface-water dynamics during this time interval.

In the modern ocean the thermocline is poised at shallow depths in the EEP, so small changes in its depth through east-west tilting can affect surface-water properties by upwelling cold, high-salinity equatorial undercurrent waters, while in the WEP the thermocline is too deep for similar-scale depth changes to induce large changes in surface-water properties. If

oscillations between El Niño- and La Niña-like climate states occurred in the equatorial Pacific on G-IG timescales, 41-kyr-paced variability in both zonal Δ SST and $\Delta\delta^{18}\text{O}_{\text{ivc-sw}}$ records should be observed with larger-scale G-IG amplitudes in SST and $\delta^{18}\text{O}_{\text{ivc-sw}}$ in the eastern compared to the western equatorial Pacific (Philander and Fedorov, 2003).

2. ENSO dynamics of the Equatorial Pacific Ocean

2.1 Modern Equatorial Pacific Ocean

In the modern tropical Pacific Ocean, zonal atmospheric circulation causes a strong asymmetry in east-to-west surface-air pressure, SST and thermocline depth along the equator (Figure 1a–c). Westward-blowing trade winds drive surface-water divergence along the equator and force warm water to pile up in the west. This results in a deep thermocline (~150 m) (Wang et al., 2000) and a thick, warm mixed layer in the WPWP and a shallow thermocline (~50 m) (Wang et al., 2000) and a thin, relatively cold and nutrient-rich mixed layer in the EEP “cold tongue”. Along the equator, modern mean annual SSTs range from ~23°C in the EEP cold tongue to ~29°C in the WPWP (Boyer et al., 2013). This ocean-atmosphere pattern is characterized by strong inter-annual variability dominated by ENSO (La Niña and El Niño events). Under La Niña conditions, a stronger atmospheric Walker Cell circulation intensifies the zonal SST gradient and tilt of the thermocline in the tropical Pacific while, during El Niño events, a weakened Walker circulation reduces this asymmetry (Figure 1b–e), resulting in more uniform temperatures across the tropical Pacific and remote effects including higher global mean temperatures, extratropical heat export and regional hydrological impacts (McPhaden et al., 2006).

2.2 Secular evolution of the equatorial Pacific climate state since the early Pliocene

An area of vigorous debate concerns the extent to which east-to-west surface-water temperature gradients along the equatorial Pacific Ocean were a feature of the geological past.

A recent study, using organic temperature proxies, suggested that a relatively constant zonal SST gradient of $\sim 3^{\circ}\text{C}$ persisted across the equatorial Pacific during the warm late Miocene and Pliocene (Zhang et al., 2014). However, Ravelo et al. (2014) posit that these data support, within error, the interpretation derived from several Mg/Ca- and model-based paleoclimate studies that the SST gradient was weak along the equator during the early Pliocene 5 to 3 Ma ago (e.g., Wara et al., 2005; Fedorov et al., 2006). A reduced zonal SST gradient in the equatorial Pacific Ocean under warm Pliocene conditions is attributed to a weak or non-existent equatorial cold tongue (Wara et al., 2005) resulting from a deep tropical thermocline relative to the modern or upwelling of warmer subsurface mode waters (Fedorov et al., 2006). This hypothesized climate state is approximated by a surface-ocean structure found today in the region during El Niño events (Wara et al., 2005; Fedorov et al., 2006) and is therefore often referred to a “permanent El Niño-like state” (e.g., Wara et al., 2005; Fedorov et al., 2015).

Regardless of whether a strong zonal SST gradient existed in the warm Pliocene equatorial Pacific Ocean (Zhang et al., 2014) or not (e.g., Wara et al., 2005; Fedorov et al., 2006), a gradual ΔSST increase since ~ 2.5 Ma, i.e., since the three major G-IG cycles of the Pleistocene (MIS 100–96) that represent the culmination of iNHG, is suggested (e.g., Wara et al., 2005; Zhang et al., 2014). This ΔSST increase is hypothesized to result from long-term deep-ocean cooling, a gradual shoaling of the tropical thermocline (Philander and Fedorov, 2003), and cooling of EEP surface-waters, while SSTs in the Western Equatorial Pacific remained relatively stable (Wara et al., 2005; Lawrence et al., 2006; Medina-Elizalde and Lea, 2010). By ~ 1.6 Ma, a Walker circulation system with a west-to-east ΔSST approximating the modern, $4\text{--}6^{\circ}\text{C}$, was reached (e.g., Wara et al., 2005; Lawrence et al., 2006).

2.3 Orbital-scale variability in equatorial Pacific surface-water structure

The results of some numerical model experiments suggest that obliquity-controlled surface-water dynamics, characterized by a see-sawing of the west-east tropical Pacific thermocline on 41-kyr periods, occurred between 3 and 1 Ma (Philander and Fedorov, 2003; Fedorov et al., 2006). These studies suggest El Niño-like and La Niña-like states associated with interglacials and glacials, respectively. Other model results suggest that the response of tropical Pacific Walker Cell circulation and thermocline tilt was not strongly influenced by changes in axial tilt during the Pleistocene (Lee and Poulsen, 2005). In line with this proposed relative insensitivity of tropical climate response to axial tilt, proxy records from the tropical Pacific are incompatible with obliquity-related changes in thermocline slope and trade-wind intensity during the early Pleistocene. For instance, calcareous nannofossil-based paleoproductivity estimates spanning MIS 101–95 demonstrate similar, almost in-phase obliquity-paced G-IG productivity pattern in the EEP and WPWP that are interpreted to be unrelated to a west-east tilting of the thermocline and thus fluctuations from an El Niño-like to a La Niña-like climate state on G-IG timescales (Bolton et al., 2010). Alternative explanations for the drivers of changes in tropical Pacific SST and productivity on G-IG (41 kyr) timescales invoke ocean-atmosphere teleconnections to southern high latitudes (Lee and Poulsen, 2005; Lawrence et al., 2006; Jakob et al., 2016) rather than local G-IG fluctuations between El Niño- and La Niña-like climate states (Philander and Fedorov, 2003; Fedorov et al., 2006).

3. Material and Methods

3.1 Site description

Here we reconstruct late Pliocene to early Pleistocene surface-water conditions in the EEP and compare them with previously published surface-water records from ODP Site 806 in the WPWP (Medina-Elizalde and Lea, 2010). Our data come from ODP Leg 138 Site 849 (0°11'N, 110°31'W; Mayer et al., 1992) (Figure 1). This site, where a nearly complete

sequence of late Miocene to Holocene sediments was recovered (Mayer et al., 1992), is situated ~860 km west of the East Pacific Rise and thus records an open-ocean signal (Mix et al., 1995). Site 849 is positioned in the present-day equatorial divergence zone within the so-called “cold tongue”. Westward-blowing trade winds induce year-round upwelling of cold, nutrient-rich equatorial undercurrent waters, resulting in mean annual SSTs at Site 849 that are up to 3°C colder compared to surface waters outside the cold tongue (Figure 1a; Boyer et al., 2013). The present-day water depth at Site 849 is 3851 m (Mix et al., 1995). This is close to the modern lysocline in the equatorial Pacific, but well above the equatorial Pacific carbonate compensation depth.

3.2 Sampling

To study planktic foraminifer-based surface-water Mg/Ca and $\delta^{18}\text{O}$ variability across iNHG in the tropical east Pacific, we analyzed 375 samples from cores 849C-7H-1-80 cm to 849C-7H-2-21 cm and 849D-6H-5-102 cm to 849D-7H-5-57 cm (77.02 to 67.78 m composite depth [mcd]). Sediments consists of white to light grey diatom nannofossil ooze, interbedded with slightly darker diatom-rich intervals (Mayer et al., 1992). Foraminifera are common within the upper 100 m (Mayer et al., 1992). For generating a record at suborbital resolution, 20 cm³ samples were taken at 2-cm intervals. Samples were oven-dried, washed over a 63 μm sieve, and weighed. Preservation of foraminiferal tests in our study interval is good, with fine features such as delicate spines and original wall structure such as pore channels well preserved, and a lack of secondary calcite on test surfaces (see Text S1 and Figure S1 in the supporting information).

To develop a high-resolution benthic foraminiferal oxygen isotope ($\delta^{18}\text{O}_\text{b}$) stratigraphy, an average of six specimens of the benthic foraminifera *Cibicidoides wuellerstorfi* were picked from the >150 μm dried sediment fraction of 125 samples from 77.02 to 74.19 mcd. The resulting data set was supplemented by data from Jakob et al. (2016)

covering the interval from 74.16 to 67.78 mcd (229 samples based on *C. wuellerstorfi*) and from Mix et al. (1995) from 76.71 to 67.71 mcd (57 samples, also based on *C. wuellerstorfi*), yielding a total of 411 stable-isotope data points for our $\delta^{18}\text{O}_b$ stratigraphy for 77.02 to 67.71 mcd.

The shallow-dwelling foraminifera *Globigerinoides ruber* (white, sensu stricto) was used to measure sea-surface $\delta^{18}\text{O}$ and Mg/Ca values across the study interval at Site 849 (77.02 to 67.78 mcd). To avoid ontogenetic effects (Elderfield et al., 2002; Friedrich et al., 2012), a narrow size fraction of 250–315 μm for $\delta^{18}\text{O}$ analyses and 200–250 μm for Mg/Ca measurements was selected. The limited number of individuals per sample within narrow size fractions dictated the use of different size fractions for $\delta^{18}\text{O}$ and Mg/Ca measurements. On average, twelve specimens were picked for $\delta^{18}\text{O}$ analyses and 40 specimens for Mg/Ca analyses. While the size fraction of *G. ruber* specimens used for $\delta^{18}\text{O}$ analyses in our study is almost identical to that used to generate the Site 806 *G. ruber*-based $\delta^{18}\text{O}$ and Mg/Ca records (250–350 μm) (Medina-Elizalde and Lea, 2010), the size fraction used for Mg/Ca is smaller in our study (200–250 μm versus 250–350 μm). Ontogenetic effects on *G. ruber*-derived Mg/Ca are a controversial issue: while Elderfield et al. (2002) record a ~ 0.5 mmol/mol increase ($\sim 2^\circ\text{C}$) from the 212–250 μm to the 300–350 μm size fraction, Friedrich et al. (2012) demonstrate no significant differences between these two size fractions.

3.3 Geochemical analyses

3.3.1 Mg/Ca analyses

For Mg/Ca measurements, tests of *G. ruber* were cracked to reveal individual chambers. The cracked shell material was carefully cleaned to remove clay minerals, organic material and re-adsorbed contaminants following the protocol of Barker et al. (2003). Cleaned samples were analyzed using the Agilent Inductively Coupled Plasma-Optical Emission Spectrometer (ICP-OES) 720 at the Institute of Earth Sciences, Heidelberg University. To allow inter-laboratory

comparison, we normalized Mg/Ca data relative to the ECRM752-1 standard applying the Mg/Ca reference value of 3.762 mmol/mol (Greaves et al., 2008). To ensure instrumental precision of the ICP-OES, the ECRM standard was analyzed at least every 20 samples. Based on replicate measurements of the ECRM reference material, we obtained a standard deviation for Mg/Ca of ± 0.02 mmol/mol (equals $\pm 0.1^\circ\text{C}$). Elemental ratios of Al/Ca, Fe/Ca and Mn/Ca were monitored to identify contamination by clay particles or diagenetic coatings. Results indicate that the Mg/Ca content of the analyzed tests is unaffected by contamination (see Text S2 and Figure S2 in the supporting information for detailed discussion).

3.3.2 Oxygen isotope analyses

Oxygen isotope analyses were performed at the Institute of Geosciences, Goethe-University Frankfurt using a ThermoFinnigan MAT253 gas source mass spectrometer equipped with a Gas Bench II. Results are reported relative to the Vienna Pee Dee Belemnite (VPDB) standard through the analysis of an in-house standard calibrated to NBS-18 and NBS-19. Analytical precision is better than 0.08‰ (at 1 σ level). The $\delta^{18}\text{O}$ values of *C. wuellerstorfi* reported herein were adjusted for the species-specific offset from equilibrium by the addition of +0.64‰ ($\delta^{18}\text{O}_{\text{adj.}}$) (Shackleton and Hall, 1984). We assume that *G. ruber* secretes calcite close to $\delta^{18}\text{O}$ equilibrium (Koutavas and Lynch-Stieglitz, 2003).

3.4 Calculation of sea-surface temperature and $\delta^{18}\text{O}_{\text{ivc-sw}}$

We converted *G. ruber* Mg/Ca (mmol/mol) into SSTs ($^\circ\text{C}$) using the species-specific equation of Lea et al. (2000) (see Text S3, Table S1 and Figure S3 in the supporting information for a detailed methodical discussion).

Ice volume-adjusted seawater $\delta^{18}\text{O}$ ($\delta^{18}\text{O}_{\text{ivc-sw}}$ in ‰ Standard Mean Ocean Water, SMOW) was estimated by subtracting an estimation of the global Plio-Pleistocene ice-volume (given by $\delta^{18}\text{O}_{\text{b}}$) from the $\delta^{18}\text{O}_{\text{sw}}$ record, following the approach of Karas et al. (2009)

(hereinafter method 1). We also used the eustatic sea-level reconstruction of Rohling et al. (2014) for the same purpose (hereinafter method 2) (see Figure S4 and Text S4 in the supporting information for a detailed methodical discussion). The zonal Pacific $\delta^{18}\text{O}_{\text{ivc-sw}}$ gradient was determined for both method 1 and 2, and, because the effect of global ice volume should be identical to both east and west Pacific records, by using the uncorrected $\delta^{18}\text{O}_{\text{sw}}$ records (hereinafter method 3).

3.5 Spectral analyses

Spectral analyses were performed to identify significant orbital frequencies in the data using the program Paleontological Statistics (PAST). To identify cyclicities and coherencies between records, Blackman-Tukey cross-spectral analyses with 30% overlap were conducted using AnalySeries. Data for cross-spectral analyses were linearly interpolated, detrended, and prewhitened; $\delta^{18}\text{O}$ values were inverted.

The relationship between Site 849 proxy records and variance at the 41-kyr period was investigated by superimposing each record on the output of a band-pass Gaussian filter centered on 0.0244 kyr^{-1} with a band width of 0.01 kyr^{-1} . The 41-kyr-filtered data were compared to the astronomical solution for obliquity (Laskar et al., 2004). Before filtering, records were detrended and linearly interpolated. Filtering was performed using AnalySeries.

3.6 Chronology

For the time interval ~2.6–2.4 Ma the Site 849 age model presented in Jakob et al. (2016) is used. A new chronology for Site 849 is established for the time interval ~2.75–2.6 Ma (Figure 2, see Table S2 in the supporting information) by visual correlation of our new $\delta^{18}\text{O}_b$ record (supplemented by data from Mix et al. [1995]) to the LR04 stack (Lisiecki and Raymo, 2005) using AnalySeries. To minimize errors during tuning, tie points at stage transitions were used (“mid-point method”) (Jakob et al., 2016, and references therein). To facilitate comparison

between Site 849 (this study) and Sites 846, 1241 and 806, we also re-tuned the benthic $\delta^{18}\text{O}$ data from these sites to the LR04 stack (Lisiecki and Raymo, 2005) across the interval ~ 2.75 – 2.4 Ma applying the “mid-point method” (see Figure S5 and Table S2 in the supporting information).

Our high-resolution $\delta^{18}\text{O}_b$ record of *C. wuellerstorfi* (Figures 2a–b, 3a) follows the 41-kyr G-IG cyclicity seen in the LR04 stack (Figure 4a, see also Figure S6 in the supporting information) and allows us to readily identify MIS G6 through 95 in our record. Oxygen-isotope values range from 2.95 to 4.31‰. For iNHG, $\delta^{18}\text{O}_b$ shows a modest long-term increase from MIS G6 to MIS 95, which is attributed to higher glacial $\delta^{18}\text{O}_b$ values (and thus more pronounced G-IG amplitudes) from MIS 100 onward.

The revised chronology for Site 849 yields a mean temporal sample resolution of ~ 800 yr and an average linear sedimentation rate of 2.7 cm kyr^{-1} (Figure 2c) for the time interval ~ 2.75 – 2.4 Ma (MIS G6–95). This is similar to the sedimentation rate of 2.6 cm kyr^{-1} in Jakob et al. (2016) for Site 849 for the time interval ~ 2.65 – 2.4 Ma (MIS G1–95).

4. Results and discussion

4.1 Plio-Pleistocene planktic foraminiferal $\delta^{18}\text{O}$ variability in the tropical Pacific

The *G. ruber*-based $\delta^{18}\text{O}$ record from Site 849 (Figure 3b) documents eight G-IG oscillations between ~ 2.75 and 2.4 Ma (MIS G6–95), closely matching variations in Earth’s obliquity (Figure 4b, see also Figure S6 in the supporting information). The lack of precessional power in planktic foraminiferal $\delta^{18}\text{O}$ and all other Site 849 records in our study may indicate that surface-water properties at that location are relatively insensitive to precessional forcing compared to G-IG obliquity-controlled climate change and associated feedbacks during the late Pliocene to early Pleistocene (see Figure S6 in the supporting information).

Planktic foraminiferal $\delta^{18}\text{O}$ values fluctuate between -1.47 and 0.11 ‰, with higher values associated with glacials and lower values associated with interglacials. Glacial-

interglacial $\delta^{18}\text{O}$ cycles increase in amplitude from MIS 100 onward ($>1\text{‰}$), with interglacial $\delta^{18}\text{O}$ values remaining invariant (around -1‰), whereas glacial $\delta^{18}\text{O}$ values shift to higher values from approximately -0.5 to 0‰ .

4.2 Plio-Pleistocene SST in the Eastern Equatorial Pacific

Our new *G. ruber* Mg/Ca-based SST record from Site 849 (Figure 3c–d) documents eight obliquity-paced G-IG oscillations between ~ 2.75 and ~ 2.4 Ma (MIS G6–95) (Figure 4c–d, see also Figure S6 in the supporting information). Considering that *G. ruber*-derived temperature estimates from the tropical Pacific most likely represent mean annual SSTs (Lin et al., 1997), our data imply mean annual SSTs ranging from $\sim 22^\circ\text{C}$ during glacial MIS 96 to $\sim 27^\circ\text{C}$ during interglacial MIS 97. Overall, our SST record documents a long-term cooling trend of $\sim 0.7^\circ\text{C}$ across the studied time interval. Glacial-interglacial SST oscillations are of larger amplitude from MIS 100 to 95 ($\sim 3\text{--}4^\circ\text{C}$) than during the time interval prior to MIS 100 ($<2^\circ\text{C}$). Comparison of our Plio-Pleistocene SST estimates with modern mean annual SSTs at Site 849 at 20 m water depth ($\sim 23.5^\circ\text{C}$; Boyer et al., 2013), i.e., the assumed mean depth habitat of *G. ruber* (Wang, 2000), indicates slightly warmer values, supporting the overall notion of a warmer-than-present EEP during the late Pliocene and early Pleistocene (e.g., Lawrence et al., 2006; Groeneveld et al., 2014).

In terms of G-IG amplitudes, our late Pliocene to early Pleistocene SST record from Site 849 is in good agreement with the pattern shown by the alkenone-derived SST record from Site 846 (Lawrence et al., 2006) and by the *Globigerinoides sacculifer* Mg/Ca-derived SST estimates from Site 1241 (Groeneveld et al., 2014) (Figure 5). In terms of absolute values, however, SSTs at Eastern Pacific Warm Pool Site 1241 appear to be substantially colder than expected when compared to modern temperatures at that location. A more southwesterly position of Site 1241 ~ 2.5 Ma ago can only account for $\sim 0.5^\circ\text{C}$ of this signal (Mix et al., 2003). The temperatures from Site 1241 are also markedly colder than SST

estimates from Sites 849 and 846, which may be related to the fact that Site 1241 *G. sacculifer*-based temperature estimates (assumed habitat at ~30 m water depth; Fairbanks et al., 1982) reflect a greater water depth compared to *G. ruber*- and alkenone-derived SSTs. Seasonality may also play a role when comparing these different proxy records. Visual evaluation of Figure 5 indicates that the Site 846 SST record leads the Site 849 SST record, suggesting, in agreement with Jakob et al. (2016), that different mechanisms dictate G-IG surface-water variability in these two distinct regions of the highly heterogeneous EEP.

4.3 Comparison between west and east tropical Pacific SST records

To explore the possibility of orbitally paced changes in surface-water structure resulting from G-IG changes in the west-east tilt of the thermocline in the Plio-Pleistocene equatorial Pacific Ocean, we evaluate zonal SST changes by comparing our *G. ruber*-based temperature record to that of Medina-Elizalde and Lea (2010) from WPWP Site 806 (see Text S5 in the supporting information for a detailed methodical discussion). Because foraminiferal preservation at Site 806 has been described as typically only moderate to good (Chaisson and Leckie, 1993; Medina-Elizalde and Lea, 2010), dissolution may have reduced foraminiferal Mg/Ca ratios, resulting in an underestimation of absolute SSTs and therefore of the reconstructed zonal SST gradient. Yet relative changes (G-IG cycles in SST at Site 806) should be preserved. However, we cannot rule out the possibility of increased and diminished carbonate dissolution during glacials and interglacials in the WEP during the early Pleistocene, respectively (Sexton and Barker, 2012). The addition of inorganic calcite from sediment pore fluids onto foraminiferal tests (overgrowths) that typically has a much larger Mg/Ca signature than its biogenic counterpart may also play a role for Site 806 *G. ruber* samples. There is, however, no correlation between Fe/Ca and Mn/Ca, diagnostic for microcrystalline overgrowth, to Mg/Ca in *G. ruber* specimens for the herein investigated time interval (Medina-Elizalde and Lea, 2010).

Sea-surface temperatures at Sites 849 and 846 in the EEP are on average $\sim 3.5^{\circ}\text{C}$ cooler than at Site 806 in the WPWP (Figures 5b, 6b). Both west and east Pacific glacial-interglacial SST (~ 41 kyr) oscillations were smaller prior to the first three large-amplitude G-IG cycles of iNHG (MIS 100–96). Glacial-interglacial SST amplitudes for MIS 100–96 are on the order of $\sim 3^{\circ}\text{C}$ in the EEP, slightly larger than the $\sim 2^{\circ}\text{C}$ amplitude in the WPWP (Figure 6b). This observation is in agreement with a previous study for the investigated time interval that compared Site 806 SSTs to east Pacific SSTs from Site 846 (Lawrence et al., 2006; Medina-Elizalde and Lea, 2010).

The zonal SST gradient between EEP Site 849 and WPWP Site 806 (Figure 6c) is relatively stable ($\sim 3^{\circ}\text{C}$) prior to MIS 102 (~ 2.64 Ma). This gradient increased progressively to $\sim 4.5^{\circ}\text{C}$ by ~ 2.4 Ma. Thus, the $\sim 4^{\circ}\text{C}$ threshold for this gradient proposed by Wara et al. (2005) based on lower-resolution records was crossed ~ 800 kyrs earlier than previously proposed, albeit with the possibility that the gradients that we reconstruct may be slight over-estimates because of the Mg/Ca size-fraction effect (see Section 3.2). Regardless, reconstructed late Pliocene to early Pleistocene west-to-east ΔSST was significantly smaller than both the late Pleistocene interglacial zonal SST gradient of $\sim 6^{\circ}\text{C}$ (which approximates modern values) (Boyer et al., 2013) and the late Pleistocene glacial gradient of $\sim 6.5^{\circ}\text{C}$ (Dyez and Ravelo, 2014). The general increase in zonal ΔSST across our study interval is mostly driven by glacial cooling in the EEP (indicated by alkenone- and planktic foraminifer-derived SST estimates; Figure 5b).

Our reconstructed zonal SST gradient, which is strongly coupled to changes in thermocline tilt (see Section 1), shows no G-IG cyclicity (Figures 6c, 7a). Therefore, our data do not support the concept of changes in thermocline slope that would signify switches between an El Niño-like and La Niña-like climate state on G-IG timescales. This finding is consistent with those of Bolton et al. (2010) for MIS 101–95 based on nannofossil assemblages.

4.4 Plio-Pleistocene $\delta^{18}\text{O}_{\text{ivc-sw}}$ at east Pacific Site 849

The five-point smoothed average of the Site 849 $\delta^{18}\text{O}_{\text{ivc-sw}}$ record (Figure 3e) indicates a clear G-IG 41-kyr cyclicity (Figure 4e), with lower $\delta^{18}\text{O}_{\text{ivc-sw}}$ during glacials particularly across prominent iNHG glacials MIS 100, 98 and 96. The amplitude of glacial-interglacial cycles in $\delta^{18}\text{O}_{\text{ivc-sw}}$ is on the order of $\sim 0.5\text{‰}$. If we assume that this residual signal represents only relative changes in sea-surface salinity (SSS), this corresponds to ~ 2.4 practical salinity units (psu) using a modern regional $\delta^{18}\text{O}_{\text{sw}}$ -SSS relationship (Schmidt et al., 1999) (see Figure S7 in the supporting information). Although a ~ 2.4 psu G-IG amplitude appears reasonable in terms of absolute values (compare Benway et al., 2006), it is important to note that this value should be only regarded as a rough estimate when applying a modern $\delta^{18}\text{O}_{\text{sw}}$ -SSS relationship to past time periods. The long-term trend of the smoothed $\delta^{18}\text{O}_{\text{ivc-sw}}$ record indicates a slight decrease over the investigated time interval of $\sim 0.1\text{‰}$, which could be interpreted as a modest freshening (~ 0.5 psu) of surface waters in the eastern tropical Pacific across iNHG (Figure 6d), although we note that such a change is within the error of the calculations.

Sea-surface salinity is generally controlled by the balance between local evaporation, local precipitation, oceanic salt advection, and fresh-water runoff. An influence of continental fresh-water runoff on SSS can be excluded at Site 849 because of its open ocean setting. It is more likely that SSS at our study site (as represented by $\delta^{18}\text{O}_{\text{ivc-sw}}$) was affected by oceanic salt advection through one or a combination of the following factors: (1) sea-level-controlled glacioeustatic closure/opening of the Central American Seaway (CAS) (Groeneveld et al., 2014), (2) changes in upwelling of high-salinity equatorial undercurrent waters (either in response to CAS closure or to changes in trade-wind strength), and (3) changes in precipitation minus evaporation.

4.4.1 Potential influence of CAS closure on Site 849

It has been suggested that glacioeustatic gateway closures of the CAS strongly influenced ocean circulation (Maier-Reimer et al., 1990; Schneider and Schmittner, 2006; Groeneveld et al., 2014). However, we suggest that the influence of temporary full CAS closure as hypothesized for MIS 100 to 98 and 96 (Groeneveld et al., 2014) on our paleoceanographic records from Site 849 can only have been minor. Reasons for this interpretation are three-fold:

(1) CAS closure would have reduced the inflow of warm, high-salinity surface waters from the Caribbean Sea into the EEP and the outflow of cold, low-salinity surface waters from the EEP into the Caribbean Sea (Maier-Reimer et al., 1990; Groeneveld et al., 2014). Thus, CAS closure should act to trap low-salinity waters west of the isthmus and expand the east Pacific fresh pool, leading to SSS variations north of the equatorial upwelling zone that are mainly controlled by CAS throughflow. If the east Pacific fresh pool expanded at least onto the equator in analogue to the modern situation as inferred from Aquarius Sea Surface Salinity images (<http://aquarius.nasa.gov>), fresher conditions during MIS 100, 99 and 98, increased salinity during MIS 97, and low-salinity conditions toward the end of MIS 96 would be expected at Site 849, based on temporary CAS closures suggested by Groeneveld et al. (2014). This pattern is, however, not seen in the SSS record from EEP Site 849.

(2) Any closure of the CAS is likely to have promoted increased equatorial upwelling rates of salinity-enriched equatorial undercurrent waters (Schneider and Schmittner, 2006). Thus, increased export production during MIS 100, 99 and 98, low productivity rates during MIS 97, and enhanced productivity toward the end of MIS 96 would be expected. However, this scenario is not supported by the productivity records from EEP Sites 849 (Jakob et al., 2016) or 846 (Lawrence et al., 2006; Bolton et al., 2010).

(3) We might expect gateway closures during MIS 100 through 98 and again toward the end of MIS 96 to result in surface-water cooling in the EEP (Groeneveld et al., 2014). However, this pattern is not consistent with SST data from east Pacific Sites 849 (this study)

and 846 (Lawrence et al., 2006). These records document substantial glacial-interglacial SST changes between MIS 100, 99 and 98 (Figure 5b), whereas, if related to CAS closure, we might also expect low SSTs during MIS 99.

4.4.2 Influence of latitudinal ITCZ migration on Site 849

For Site 849, upwelling intensification, diagnostic of increased easterly trade-wind strength, has been demonstrated from MIS 100 onward, with peak upwelling occurring toward glacial terminations (Jakob et al., 2016). We might therefore reasonably expect increased SSSs through the influence of high-salinity equatorial undercurrent waters. Yet our data record lower-salinity surface waters for glacials MIS 100, 98 and 96, superimposed on a modest long-term freshening of surface waters in the EEP from MIS G6 through 95 (Figure 3e). We suggest that enhanced precipitation during glacials relative to interglacials is responsible for this pattern. This hypothesis invokes an equatorward migration of the Intertropical Convergence Zone (ITCZ) during glacials in the eastern Pacific. The modern ITCZ shows strong seasonal meridional migration toward the warmer hemisphere in response to insolation forcing (Figure 1). Paradoxically, however, the ITCZ migrates toward the cooler hemisphere in response to ENSO dynamics on inter-annual timescales, primarily driven by tropical changes in atmospheric energy input (Schneider et al., 2014, and references therein) (see Section 4.7). Latitudinal ITCZ migration is also inferred from paleo-records in response to changes in inter-hemispheric temperature gradients on millennial through G-IG and secular timescales (e.g., Koutavas and Lynch-Stieglitz, 2003; Benway et al., 2006; Schneider et al., 2014; Hyeong et al., 2016). Strong cooling in the high-northern latitudes during prominent iNHG glacials (Lawrence et al., 2009) is a plausible mechanism for equatorward (i.e., southward) movement of the ITCZ. The amplitude of suggested G-IG paleosalinity change (~2.4 psu) at Site 849 (Figure 3e) in response to ITCZ shifts in the eastern tropical Pacific during iNHG falls in the range of those estimated for late Pleistocene ITCZ-controlled

paleosalinity variations of ~2–4 psu (Benway et al., 2006). In light of the strong control exerted by Atlantic Meridional Overturning Circulation on the strength of the global annual mean cross-equatorial temperature gradient, an ITCZ influence on our records is consistent with evidence for major southern component water incursions into the deep North Atlantic (implied Atlantic Meridional Overturning Circulation weakening or shoaling) during MIS G6, 100 and 96 (Lang et al., 2016).

4.5 Possible influences on the paleosalinity record from the WPWP

The $\delta^{18}\text{O}_{\text{ivc-sw}}$ record from Site 806 in the WPWP varies on 41-kyr timescales (Figure 6d). If most of this signal represents changes in local salinity, the Site 806 $\delta^{18}\text{O}_{\text{ivc-sw}}$ record indicates fresher conditions during glacials than during interglacials (Figure 6d). This is similar to the observation derived from Site 849 in the EEP for prominent iNHG glacial-interglacial cycles. Modeling and proxy data consistently indicate lower-salinity glacial conditions for the Last Glacial Maximum for both the WPWP (e.g., Lea et al., 2000; DiNezio and Tierney, 2013) and the EEP cold tongue (Koutavas and Lynch-Stieglitz, 2003), consistent with our observations for early Pleistocene glacials.

As discussed in Section 4.4.2, a southward movement of the ITCZ during glacials may have promoted an increasing influence of the ITCZ rain belt on east Pacific Site 849. A contemporaneous southward displacement of the mean ITCZ in the west Pacific during glacials to a position akin to that observed during modern winter conditions, however, would have prevented a year-round influence of the ITCZ on Site 806 (Figure 1). Hence, saltier water would be expected in the WEP during glacials than during interglacials. This scenario is, however, not compatible with the local Site 806 $\delta^{18}\text{O}_{\text{ivc-sw}}$ record (Medina-Elizalde and Lea, 2010), which indicates fresher conditions during glacials than during interglacials (Figure 6d). There are many different possible explanations for this result, including, for instance, a different Plio-/Pleistocene pathway of ITCZ migration in the WEP from today.

Additionally, it might be speculated that a glacioeustatically driven exposure of the Sunda Shelf may have influenced the Plio-Pleistocene paleosalinity pattern in the WPWP. The exposure of the Sunda Shelf is an important factor influencing regional glacial climates (Sun et al., 2000; DiNezio and Tierney, 2013). Pollen assemblages from the continental shelf of the South China Sea document intensified river flows on the exposed Sunda Shelf in response to strengthened winter monsoon precipitation during the Last Glacial Maximum (Sun et al., 2000). Some model results for the Last Glacial Maximum are consistent with this observation, indicating that a change in the Walker circulation driven by weakened atmospheric convection over the WPWP results in (i) increased riverine freshwater flux in response to enhanced precipitation over the exposed land and (ii) increased rainfall in the west Pacific region, thereby reducing SSS over much of the western tropical Pacific (DiNezio and Tierney, 2013). Given the distance between the exposed Sunda Shelf and Site 806 (~5500 km), the latter mechanism is the more plausible of these two explanations for SSS change at Site 806. We note, however, that this interpretation of hydroclimate in the WPWP is contradicted by some other model results summarized in DiNezio and Tierney (2013) and by speleothem records from northern Borneo indicating drier conditions during late Pleistocene glacials rather than during interglacials (Meckler et al., 2012).

4.6 Comparison between west and east tropical Pacific $\delta^{18}\text{O}_{\text{ivc-sw}}$ records

To provide further clues to potential ITCZ migrations and the apparent lack of glacial-interglacial thermocline-tilt-related changes in SSTs in the equatorial Pacific during the Plio-Pleistocene, we evaluate west-to-east $\delta^{18}\text{O}_{\text{ivc-sw}}$ gradients (see Text S5 in the supporting information for detailed methodical discussion).

4.6.1 Robustness of approaches

The three different approaches that we have applied to adjust our records for changes in continental ice volume (see Section 3.4) yield identical results in terms of G-IG amplitudes and trends of the zonal equatorial Pacific $\delta^{18}\text{O}_{\text{ivc-sw}}$ ($\Delta\delta^{18}\text{O}_{\text{ivc-sw}}$) and $\delta^{18}\text{O}_{\text{sw}}$ ($\Delta\delta^{18}\text{O}_{\text{sw}}$), gradient (Figure 6e–f). This observation suggests that this result is robust (albeit lower absolute values are obtained using method 3). For simplification, we refer to method 1 only in the follow-up discussion.

4.6.2 Obliquity oscillations in the zonal Pacific $\delta^{18}\text{O}_{\text{ivc-sw}}$ gradient

The zonal Pacific $\delta^{18}\text{O}_{\text{ivc-sw}}$ gradient (interpreted to represent changes in the zonal SSS gradient; see Sections 4.4 and 4.5) documents saltier surface water by ~2.4 psu (as defined by ~0.5‰) in the EEP than in the WPWP at ~2.75 Ma (Figure 6e). Over the studied time interval, the long-term salinity gradient decreased, disappeared around MIS 100, and increased again from MIS 100 onward.

Our new $\Delta\delta^{18}\text{O}_{\text{ivc-sw}}$ record provides strong evidence for the existence of ~41-kyr fluctuations in phase with global ice-volume during the entire study interval (Figure 7b). If these 41-kyr oscillations were a result of changes in thermocline tilt, we might expect higher-amplitude G-IG surface-water salinity variations in the EEP than in the WPWP (see Section 1). However, G-IG surface-water salinity variability of similar magnitude in the Eastern and Western Equatorial Pacific during the most prominent iNHG glacial-interglacial cycles, as evidenced by our data, argues against thermocline-tilt-controlled surface-water dynamics (Figure 6d). This result supports our hypothesis that G-IG salinity changes in the tropical Pacific were more likely driven by ITCZ displacement over the EEP. In combination with our orbital-resolution zonal ΔSST record (Figure 7a), these findings have important implications for the ongoing debate on the existence of obliquity-paced oscillations from El Niño- to La Niña-like climate states in the tropical Pacific during the Plio-Pleistocene, which we conclude were unlikely during late Pliocene to early Pleistocene G-IG cycles.

4.7 Implications for obliquity-paced surface-water dynamics during the Plio-Pleistocene

The typical pattern of seasonal ITCZ migration toward the warmer hemisphere caused by insolation changes is not observed during modern ENSO variability (Schneider et al., 2014). Going from La Niña to El Niño conditions in the modern, the northern hemisphere warms more than the southern hemisphere. Paradoxically, the ITCZ responds with a southward migration toward the cooler hemisphere, which appears to be primarily driven by tropical changes in atmospheric energy input (Schneider et al., 2014, and references therein). This has important implications for the debate over whether surface-water dynamics in the equatorial Pacific followed glacial-interglacial-scale ENSO-like pattern across iNHG. We interpret local $\delta^{18}\text{O}_{\text{ivc-sw}}$ changes in the late Pliocene to early Pleistocene EEP to indicate a northward migration of the ITCZ during interglacials and a southward migration during glacials. This interpretation is supported by both our zonal equatorial Pacific ΔSST (which are unrelated to G-IG cycles; Figure 7a) and $\Delta\delta^{18}\text{O}_{\text{ivc-sw}}$ estimates (which show a G-IG cyclicity, but suggest an ITCZ movement toward the warmer hemisphere during prominent iNHG glacial-interglacial cycles; Figure 7b). Evaluation of the applicability of modern atmospheric ENSO and ITCZ dynamics to these longer G-IG timescales is, however, required.

5. Conclusions

We present new high-resolution records from Eastern Equatorial Pacific ODP Site 849 and compare them to west Pacific ODP Site 806 to critically assess the possibility of obliquity-paced changes in surface-water structure, involving changes in Walker Cell circulation, during the Plio-Pleistocene onset of Northern Hemisphere Glaciation (MIS G6–95, ~2.75–2.4 Ma). The most salient features of our study are as follows:

(1) We find that the west-east equatorial Pacific SST gradient progressively increased across iNHG, mostly driven by glacial cooling in the EEP. The absence of a G-IG cyclicity in

the zonal equatorial Pacific SST gradient argues against obliquity-paced fluctuations from El Niño-like to La Niña-like climate states in the equatorial Pacific during the late Pliocene and early Pleistocene.

(2) In analogy to modern seasonal ITCZ shifts, we propose the reconstructed glacial-interglacial SSS amplitudes (as represented by $\delta^{18}\text{O}_{\text{ivc-sw}}$) at Site 849 during most prominent iNHG glacial-interglacial cycles to be related to a glacial southward and interglacial northward displacement of the ITCZ rain belt, respectively.

(3) The equatorial Pacific SSS gradient (as represented by $\Delta\delta^{18}\text{O}_{\text{ivc-sw}}$) is characterized by 41-kyr fluctuations. We speculate that glacial-interglacial north-south migrations of the ITCZ over the EEP as evidenced in the $\delta^{18}\text{O}_{\text{ivc-sw}}$ record from Site 849 controlled the $\delta^{18}\text{O}_{\text{ivc-sw}}$ gradient in the Pacific during prominent iNHG glacials. These findings have important implications for the debate on possible fluctuations between El Niño-like and La Niña-like climate states on G-IG timescales, which we conclude were unlikely during the late Pliocene and early Pleistocene.

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Figure captions

Figure 1: Location map of ODP Sites 849 (blue, this study; 0°11’N, 110°31’W, 3851 m [Mayer et al., 1992]), 846 (black, 3°5’N, 90°49’W, 3296 m), 806 (black, 0°19’N, 159°21’E, 2520 m) and 1241 (black, 5°51’N, 86°27’W, 2027 m). Modern mean (a) annual, (b) boreal summer (Jul–Sep) and (c) boreal winter (Jan–Mar) SSTs (°C), and modern mean (d) boreal summer (Jul–Sep) and (e) boreal winter (Jan–Mar) SSSs (psu) are shown after World Ocean Atlas (Boyer et al., 2013). Summer SSTs approximate modern La Niña conditions, while winter SSTs resemble modern El Niño conditions. White lines show the modern northernmost

and southernmost positions of the Intertropical Convergence Zone (ITCZ) in July and January, respectively. Modern ITCZ positions were determined using 55-year (1958–2013) satellite observations of monthly rainfall in the tropical Pacific provided by the NASA Tropical Rainfall Measuring Mission (<http://trmm.gsfc.nasa.gov>).

Figure 2: Revised age model, tuned to the LR04 stack (Lisiecki and Raymo, 2005), for ODP Site 849 spanning MIS G6–95 (~2.75–2.4 Ma). (a) LR04 stack for MIS G6–95 (Lisiecki and Raymo, 2005). (b) Benthic foraminiferal (*C. wuellerstorfi*) $\delta^{18}\text{O}$ record from Site 849. The record consists of new data supplemented by published data of Mix et al. (1995) and Jakob et al. (2016). Bars indicate the 1σ standard deviation. Blue dashed lines in Figure 2a–b show the tie points used for tuning. Tuning points are provided in Table S2. (c) Average linear sedimentation rate (LSR) in cm kyr^{-1} . Grey bars highlight glacial periods.

Figure 3: Planktic foraminiferal temperature and salinity proxy data based on our revised benthic foraminiferal stratigraphy from ODP Site 849 for MIS G6–95 (~2.75–2.4 Ma). (a) Benthic foraminiferal (*C. wuellerstorfi*) $\delta^{18}\text{O}$ stratigraphy. (b) Planktic foraminiferal (*G. ruber*) $\delta^{18}\text{O}$ record ($\delta^{18}\text{O}_p$). (c) *Globigerinoides ruber* Mg/Ca record adjusted to the ECRM standard (Greaves et al., 2008). (d) *Globigerinoides ruber* Mg/Ca-based SST estimates. (e) Ice-volume-adjusted $\delta^{18}\text{O}_{\text{sw}}$ record ($\delta^{18}\text{O}_{\text{ivc-sw}}$) as a proxy for SSS. Higher (lower) $\delta^{18}\text{O}_{\text{ivc-sw}}$ values are interpreted to indicate saltier (fresher) sea-surface waters at Site 849. Using a regional $\delta^{18}\text{O}_{\text{sw}}$ –SSS relationships, 1 psu corresponds to ~0.2‰ (Schmidt et al., 1999). Grey points in Figure 3d–e depict individual data points, black line illustrates the five-point smoothed average of a record. Horizontal and vertical bars indicate the 1σ standard deviation associated with the age model and the individual proxy records, respectively. Grey bars highlight glacial periods.

777

778 **Figure 4:** 41-kyr-filtered data (red) for Site 849 proxy records for ~2.75–2.4 Ma (MIS G6–
779 95) compared to the astronomical solution for orbital obliquity (black; Laskar et al., 2004) for
780 (a) benthic foraminiferal (*C. wuellerstorfi*) $\delta^{18}\text{O}$, (b) planktic foraminiferal (*G. ruber*) $\delta^{18}\text{O}$
781 ($\delta^{18}\text{O}_p$), (c) *G. ruber* Mg/Ca, (d) *G. ruber* Mg/Ca-based SST estimates, (e) ice-volume-
782 adjusted $\delta^{18}\text{O}_{sw}$. Grey points in Figure 4d–e depict individual data points, black line illustrates
783 the five-point smoothed average of a record. Grey bars highlight glacial periods.

784

785 **Figure 5:** East Pacific SST estimates from ODP Sites 849 (black; five-point smoothed; this
786 study) and 846 (red; Lawrence et al., 2006) for MIS G6–95 (~2.75–2.4 Ma) and from ODP
787 Site 1241 (purple; Groeneveld et al., 2014) for MIS 100–95 (~2.55–2.42 Ma). Records from
788 Sites 846 and 1241 are re-tuned to LR04 (Lisiecki and Raymo, 2005), tuning points are
789 provided in Table S2. (a) Benthic foraminiferal (*C. wuellerstorfi*) $\delta^{18}\text{O}$ stratigraphy for Site
790 849. (b) East Pacific SSTs. While estimates from Site 849 (this study; Mg/Ca_{*G. ruber*}-based)
791 and 846 (Lawrence et al., 2006; alkenone-based [C₃₇]) reflect temperatures of the uppermost
792 part of the water column, temperature estimates from Site 1241 (Groeneveld et al., 2014;
793 Mg/Ca_{*G. sacculifer*}-based) reflect a water depth of ~30 m (Fairbanks et al., 1982). Thereafter,
794 black, red and purple arrows represent modern mean annual SSTs at Sites 846 (surface water),
795 849 (20 m water depth) and 1241 (30 m water depth), respectively (Boyer et al., 2013).
796 Dashed lines indicate trends. Horizontal and vertical bars indicate the 1 σ standard deviation
797 associated with the age model and the individual Site 849 proxy records, respectively. Grey
798 bars highlight glacial periods.

799

800 **Figure 6:** High-resolution planktic foraminiferal temperature and salinity proxy data from
801 ODP Site 849 (black; this study) in the EEP and ODP Site 806 (blue; Medina-Elizalde and
802 Lea, 2010) in the WEP based on revised benthic foraminiferal stratigraphies from MIS G6–95

803 (~2.75–2.4 Ma) (tuning points are provided in Table S2). (a) Benthic foraminiferal (*C.*
 804 *wuellerstorfi*) $\delta^{18}\text{O}$ stratigraphy for Site 849. (b) *Globigerinoides ruber* Mg/Ca-based SST
 805 estimates from Sites 849 in the EEP (black; five-point smoothed; this study) and 806 in the
 806 WEP (blue; Medina-Elizalde and Lea, 2010). Black and blue arrows represent modern mean
 807 annual SSTs at 20 m water depth for Sites 849 and 806, respectively (Boyer et al., 2013). (c)
 808 Difference between *G. ruber* Mg/Ca-based SST estimates from Sites 849 in the EEP (this
 809 study) and 806 in the WEP (Medina-Elizalde and Lea, 2010) ($\Delta\text{SST}_{\text{west-east}}$). Pale purple
 810 shaded zone represents the range of the modern (~6°C; Boyer et al., 2013) to late Pleistocene
 811 (~4°C; Wara et al., 2005) SST gradients between the EEP and WEP. (d) Planktic
 812 foraminiferal (*G. ruber*) $\delta^{18}\text{O}_{\text{ivc-sw}}$ record from Sites 849 in the EEP (black; five-point
 813 smoothed; this study) and 806 in the WEP (blue; Medina-Elizalde and Lea, 2010). (e)
 814 Difference between $\delta^{18}\text{O}_{\text{ivc-sw}}$ from Sites 849 in the EEP (this study) and 806 in the WEP
 815 (Medina-Elizalde and Lea, 2010) following method 1 (purple) and method 2 (pink). Pale
 816 purple shaded zone represents the modern $\delta^{18}\text{O}_{\text{ivc-sw}}$ gradient between the EEP and the WEP
 817 (Schmidt et al., 1999). (f) Difference between $\delta^{18}\text{O}_{\text{sw}}$ from Sites 849 in the EEP (this study)
 818 and 806 in the WEP (Medina-Elizalde and Lea, 2010) following method 3 (grey). The break
 819 at ~2450 ka represents a coring gap in the Site 806 record. Dashed lines indicate trends.
 820 Horizontal and vertical bars indicate the 1σ standard deviation associated with the age model
 821 and the individual Site 849 proxy records, respectively. Grey bars highlight glacial periods.

822
 823 **Figure 7:** (top) Spectral analyses of (a) Site 849 $\delta^{18}\text{O}_b$ (purple) and tropical Pacific zonal
 824 ΔSST (dark blue) and of (b) Site 849 $\delta^{18}\text{O}_b$ (purple) and tropical Pacific zonal $\Delta\delta^{18}\text{O}_{\text{ivc-sw}}$
 825 (method 1) (dark blue) for the time interval ~2.75–2.4 Ma. The highest confidence level is
 826 shown for each record (dashed lines). (bottom) Blackman-Tukey cross-spectral analyses
 827 showing coherence (red) and phase (grey) relationships between global ice volume (Site 849,
 828 $\delta^{18}\text{O}_b$) and the tropical Pacific zonal (a) SST and (b) $\delta^{18}\text{O}_{\text{ivc-sw}}$ gradient (method 1) between

829 Sites 849 in the EEP and 806 in the WEP spanning the time interval ~2.75–2.4 Ma. Black
830 dashed lines mark 95% and 80% confidence levels. Blue lines indicate obliquity (41 kyr),
831 precessional (23 kyr, 19 kyr) and half-precessional (11 kyr) cycles. Note log scale on x-axes.
832 The $\delta^{18}\text{O}_b$ - ΔSST cross-spectral analysis indicates no coherency at the 41-kyr band, indicating
833 that the west-to-east Pacific SST gradient is unrelated to obliquity. The existence of 41-kyr
834 cycles in the individual records for Site 849 and 806 $\delta^{18}\text{O}_{\text{ivc-sw}}$ calculation is evident in the
835 single spectrum analyses, each showing significant peaks at the 41-kyr band (see Figure S6 in
836 the supporting information).

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Table 1

ODP Site 849

Age	graphical correlation of our new benthic foraminiferal $\delta^{18}\text{O}$
LSR	linear sedimentation rate
FFI	planktic foraminiferal fragmentation index (supplemer
$\delta^{18}\text{O}$ <i>C. wuellerstorfi</i>	adjusted to equilibrium calcite by the addition of +0.64
Mg/Ca <i>G. ruber</i>	reported Mg/Ca values are normalized to the ECRM752
SST <i>G. ruber</i>	Mg/Ca based SST estimates following Lea et al. (2000).
$\delta^{18}\text{O}$ sw	calculated using the high-light paleotemperature equat
$\delta^{18}\text{O}$ sw-ivc	$\delta^{18}\text{O}$ sw, corrected for ice volume
*)	data from Mix et al. (1995)

Age (ka)	Depth (mcd)	Site	Hole	Core
2385.57	67.71	849	D	6
2387.85	67.78	849	D	6
2388.83	67.81	849	D	6
2389.49	67.83	849	D	6
2389.81	67.84	849	D	6
2391.45	67.89	849	D	6
2392.10	67.91	849	D	6
2394.06	67.97	849	D	6
2394.39	67.98	849	D	6
2395.37	68.01	849	D	6
2396.02	68.03	849	D	6
2396.67	68.05	849	D	6
2397.16	68.06	849	D	6
2397.65	68.08	849	D	6
2398.31	68.10	849	D	6
2398.64	68.11	849	D	6
2398.96	68.12	849	D	6
2399.94	68.15	849	D	6
2400.60	68.17	849	D	6
2401.26	68.19	849	D	6
2401.91	68.21	849	D	6
2403.88	68.27	849	D	6
2404.54	68.29	849	D	6
2405.20	68.31	849	D	6
2405.53	68.32	849	D	6
2407.84	68.39	849	D	6

Table 2

Age
δ18O sw-ivc
ΔSST
Δδ18O sw-ivc (method 1)
Δδ18O sw-ivc (method 2)
Δδ18O sw (method 3)

graphical correlation of benthic foraminiferal δ18O
δ18O sw for Site 806 (Medina-Elizalde and Lea, 201
SST gradient between Site 806 and 849 were calcul
gradient between Site 806 and 849 based on δ18Os
gradient between Site 806 and 849 based on δ18Os
gradient between Site 806 and 849 δ18Osw

Age	d ¹⁸ Osw-ivc 806	ΔSST 806-849	Δδ18O sw-ivc 806-849 (‰ SMOW)
(ka)	(‰ SMOW)	(°C)	(method 1)
2388.20	1.62	3.63	0.34
2390.44	1.57	3.83	0.29
2392.51	1.49	3.83	0.23
2395.54	1.74	4.40	0.48
2398.60	1.40	3.76	0.12
2401.15	1.30	3.77	0.00
2404.43	1.31	3.67	-0.03
2407.51	1.33	3.48	0.00
2409.66	1.37	3.40	-0.02
2412.08	1.80	3.80	0.36
2414.38		3.13	
2415.96	1.85	3.52	0.31
2417.82	1.66	2.92	0.12
2419.37	1.65	3.42	0.04
2420.58	1.64	3.75	0.14
2421.80	1.68	3.62	0.21
2422.94	1.88	4.16	0.42
2423.72	1.96	4.03	0.57
2424.61	2.05	4.53	0.73
2425.22	1.90	4.19	0.48
2425.93	2.10	5.05	0.62
2426.54	2.04	5.46	0.55
2426.99	1.96	5.40	0.47
2427.54	1.76	4.30	0.31
2428.07	1.79	4.54	0.45
2428.48	1.23	4.09	0.10
2428.98	1.52	5.38	0.57
2429.55		4.80	
2430.00	1.22	4.37	0.35
2430.64	1.04	5.15	0.19
2431.30	0.97	4.92	0.10