Evaluating drivers of Pleistocene eastern tropical Pacific sea surface temperature

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Abstract Sea surface temperature (SST) of the eastern equatorial Pacific is a key component of tropical oceanic and atmospheric circulation with global teleconnections. Forcing factors such as local and high-latitude insolation changes, ice sheet size and albedo feedbacks, and greenhouse gas radiation have been proposed as controls of long-term tropical Pacific SST, though the precise role each mechanism plays is not fully known on glacial-interglacial or longer timescales. Here proposed mechanisms are evaluated by comparing orbital-scale records of eastern Pacific SST with forcing variability over the past 1.5 Ma. The primary SST records are a compilation of new and existing data from Ocean Drilling Program Site 1239 at the northeastern margin of the modern eastern Pacific cold tongue and Site 846 SST within the cold tongue. Using time series analysis, we test previously proposed mechanisms for control of long-term tropical SST change and SST gradients in the eastern Pacific. We find that within statistical uncertainties, in the precession band early eastern Pacific SST is consistent with direct forcing by equatorial radiation changes in the tropical cold season (summer-fall) rather than inversely correlated as previously suggested. In the obliquity band high-latitude solar forcing leads or is in phase with eastern equatorial Pacific SST, while in the eccentricity band atmospheric greenhouse gas concentrations are closely associated with cold tongue SST. Pleistocene eastern Pacific SST gradients indicate that the gradient on the northern margin of the cold tongue strengthened through the mid-Pleistocene transition, a result compatible with the cold tongue becoming more focused at ~900–650 ka.

1. Introduction

Changes in eastern equatorial Pacific (EEP) sea surface temperature (SST) impact regional pressure gradients and consequently atmospheric and ocean circulation [e.g., Cane and Clement, 1999; Clement and Cane, 1999]. In the early Pleistocene, the SST record from Site 846 in the heart of the EEP cold tongue indicates long-term cooling, in both the minimum glacial and maximum interglacial temperatures, especially from 1.4 to 0.9 Ma, with variations on orbital timescales (tens of thousands of years) superimposed on this longer-term trend. Orbital-scale EEP SST adjustments are proposed to have been the result of such factors as high-latitude and/or equatorial insolation changes [e.g., Hays et al., 1976; Koutavas et al., 2002; Philander and Fedorov, 2003; Timmermann et al., 2007; Martínez-García et al., 2010], climate dynamics related to ice sheets [e.g., Imbrie et al., 1992], and/or atmospheric $p$CO$_2$ changes [e.g., Lea et al., 2000]. These factors are not independent of one another—indeed, many share common elements, such as their influence on ice sheets or on the state of the ventilated thermocline. However, through time series coherence and phasing relationships, it may be possible in most cases to identify the primary source of EEP SST forcing at certain frequencies in the Pleistocene. These mechanisms have been previously proposed for shorter timescales [e.g., Spero and Lea, 2002; Pena et al., 2008], and here we examine their activity over the past 1.5 Ma.

Records of EEP SST can be divided at the mid-Pleistocene transition (MPT, ~1.0–0.8 Ma), a period when Pleistocene marine $\delta^{18}O$ records progressively shifted from being dominated by oscillations at the 41 kyr periodicity to a more asymmetric longer-term rhythm with colder glacialis; this climate shift is likely not the result of a regime change in orbital insolation forcing but rather either the gradual movement over a threshold that changed internal Earth system feedbacks or a more abrupt climate event [e.g., Pisias and Moore, 1981; Elkibbi and Rial, 2001; Clark et al., 2006; Kitamura, 2015]. In light of the lack of an obvious insolation-related mechanism for this MPT shift, possible Earth system explanations include changes in the behavior of ice sheets [Imbrie et al., 1993; Clark and Pollard, 1998; Gildor and Tziperman, 2000, 2001; Clark et al., 2006; McClymont et al., 2008] and/or periodic oceanic carbon sequestration and release [e.g., Schmieder et al., 2000; Toggweiler et al., 2006;...
Watson and Naveira Garabato, 2006; Kemp et al., 2010; Sigman et al., 2010; Pena and Goldstein, 2014; Maslin and Brierley, 2015]. Tropical ocean dynamics could play a role in either of these factors; for instance, cooling tropical Pacific SST prior to 0.9 Ma could be linked to a glacial pCO2 decline or could have been a precondition for the expansion of ice sheets [McClymont et al., 2013]. In this manuscript we examine four potential drivers of climate: (1) local equatorial insolation in October, (2) high-latitude orbital forcing, (3) ice sheet forcing, and (4) CO2 forcing. These forcing mechanisms are considered independently for the periods immediately preceding and after the MPT.

The primary objective of this work is to evaluate existing hypotheses for variations in eastern tropical Pacific SST prior to and following the MPT. We use time series coherence and phase lag testing for this evaluation. Long, continuous, and orbitally resolved records of tropical SST are essential to this approach, one of which is derived from the heart of the EEP cold tongue at Ocean Drilling Program (ODP) Site 846 [Liu and Herbert, 2004; Lawrence et al., 2006]. A counterpart to this eastern Pacific SST record is available from the western Pacific warm pool at ODP Site 806 [Medina-Elizalde and Lea, 2005], though here we chiefly focus on the EEP. To better constrain the development and variations of SST in the EEP cold tongue, we generated additional alkenone-based SST data from the Ecuador margin at ODP Site 1239 and compared this record with other tropical localities to test proposed mechanisms for control of Pleistocene SST patterns in the tropical Pacific. Below, we first review possible mechanisms for changing SST in the EEP and then use new and existing data to test for concordant behavior with these hypotheses.

2. Hypotheses for EEP Temperature Change
2.1. Direct Local Solar Forcing of the EEP Cold Tongue
Direct equatorial insolation forcing has been proposed to have controlled cold tongue SST [Clement et al., 1999, 2000; Koutavas et al., 2002], though it may not be immediately obvious which season, if any, should dominate cold tongue SST. Seasonal insolation at the equator varies by as much as 60 W/m² in the precessional (23 kyr) band; numerical models suggest that such insolation in the boreal autumn has the potential to alter tropical Pacific west-east (zonal) and north-south (meridional) gradients through asymmetric heating of the eastern and western portions of the equatorial Pacific [e.g., Clement et al., 1999]. The effect is suggested to be greatest in the eastern Pacific where the Intertropical Convergence Zone (ITCZ) is more tightly constrained and north of the equator (Figure 1) due to interhemispheric differences in landmass geometry and distribution [Philander et al., 1996]. In boreal winter and spring, the ITCZ roughly lies along the equator in both the eastern and western Pacific; hence, changes in winter/spring equatorial insolation, which are uniform along the equator, do not greatly modify the zonal surface temperature gradient [e.g., Pisas and Moore, 1981; Clement et al., 1999; Elkibbi and Rial, 2001]. However, in the boreal late summer and early autumn, the modern ITCZ shifts north of the equator in the east but remains close to the equator in the west Pacific. The more northerly position of the ITCZ in the eastern tropical Pacific is linked to strong divergent equatorial trade winds and cool-water upwelling at the equator, especially near the Galapagos Islands, thereby resulting in the strong eastern equatorial cold tongue. The contrast between the equatorial cold tongue and the warmer water both in the western Pacific and the eastern Pacific warm pools may further enhance upwelling and cool the cold tongue. Temporal changes in equatorial solar heating may have the strongest impact during the time of year when increased insolation would warm the west more than the east (Northern
Hemisphere (NH) autumn), thereby strengthening the west-east SST gradient [Mitchell and Wallace, 1992]. The increased zonal SST gradient and stronger Walker circulation feedback could lead to intensified upwelling in the eastern Pacific cold tongue and could reinforce the position of warm atmospheric convergence north of the equator in the eastern Pacific. As a result, periods of increased boreal summer-fall equatorial insolation could combine with the ITZ and pressure gradient feedbacks to cool temperatures in the cold tongue [Clement et al., 1999; Koutavas et al., 2002]. The hypothesis that the strength of October insolation controlled orbital-scale EEP SST would be validated by finding a coherent and in-phase relationship in a spectral comparison of SST records with calculated October insolation values. While other forcing factors (ice sheet changes and CO₂ concentrations) could introduce threshold responses and trigger longer-term change over hundreds of thousands of years, the frequency of equatorial insolation has not changed over this time period (1.5 Ma to present) and is therefore less likely to introduce these longer-term changes in SST.

2.2. Remote High-Latitude Solar Forcing

Variations in eastern Pacific SST may also be influenced by insolation variability outside the tropics through ocean circulation [e.g., Philander and Fedorov, 2003; Martínez-Garcia et al., 2010]. In the higher-latitude Pacific, the surface ocean loses heat to the atmosphere, and cool surface water subducts to form shallow mode water that travels equatorward along the subtropical thermocline, reaching the surface again through upwelling in the eastern Pacific [Nakamura, 1996; Suga et al., 1997] where it is reheated in the tropics. This process occurs in both the Northern and Southern Hemispheres, though the exact transport pathways are not symmetric in the two hemispheres [Harper, 2000]. Heat loss at higher latitudes is thus balanced by heat gain at low latitudes at decadal-to-centennial timescales [Philander and Fedorov, 2003].

High-latitude insolation forcing, if it alters the conditions where mode water subducts, may thus impact the ventilated thermocline and therefore tropical regions where thermocline water upwells. Although either changes in Earth’s tilt (obliquity) or the precession of the equinoxes alter high-latitude insolation, we focus here on obliquity. Only changes in tilt modify the annual average latitudinal distribution of insolation, whereas changes in precession primarily affect the seasonal variations. As obliquity increases, the latitudinal surface energy gradient and subsurface density gradients in subpolar regions also increase in the Northern or Southern Hemisphere [Rubincam, 1994; Philander and Fedorov, 2003]. We test for a connection between insolation cycles, thermocline conditions, and low-latitude SST by comparing high-latitude (e.g., 65°N) insolation curves and tropical SST records. If high-latitude obliquity forcing controls EEP SST, we should find these records to be coherent and in phase on orbital timescales. If a connection exists, high-latitude insolation should also be coherent and in phase with the changes in the west-east equatorial SST gradient as thermocline variability could be expected to have a greater influence on SST in the eastern Pacific where the thermocline is closer to the surface.

2.3. Remote Ice Sheet Forcing

Although variations in the buildup and reduction of ice sheets strongly influence high-latitude climate, they also potentially affect atmospheric circulation and SST in remote areas such as the tropical Pacific, primarily through changes in orography [Chiang, 2003; Chiang and Bitz, 2005; Ruddiman, 2006]. For example, large Northern Hemisphere ice sheets may have displaced the mean latitude of atmospheric bands—including the ITZ and equatorial front—southward of present positions during glacial intervals [Chiang and Bitz, 2005; Koutavas and Lynch-Stieglitz, 2005; Timmermann et al., 2007; Rincón-Martínez et al., 2010; Toggweiler and Lea, 2010]. This being the case, ice sheets are not, however, independent from other mechanisms of long-term climate change; high-latitude solar forcing and radiative greenhouse gas concentrations, listed here as alternative mechanisms, both have the potential to induce changes in ice sheet thickness and extent. Coherency, amplitude, and phase lag relationships allow for tests of the potential effects of the buildup and melting of ice sheets on tropical temperatures. Ice sheet thickness (orography) and areal extent (albedo) may well have different forcing effects on climate. If Laurentide ice sheet orography changed EEP SST by shifting atmospheric zones southward after the MPT, benthic δ¹⁸O (as an approximate ice volume proxy, but see discussion in section 5.4) should be coherent and in phase with EEP SST at orbital timescales. Intuitively, ice sheet orography might be expected to be a more important factor after the MPT if ice sheets became thicker at that time.
2.4. Greenhouse Gas Forcing

Tropical SST changes over the past 1.5 Ma may also be related to variations in radiative forcing due to atmospheric greenhouse gas (GHG) concentrations [Lea, 2004; Medina-Elizalde and Lea, 2005; Lea et al., 2006] which have been suggested to contribute roughly one third to two thirds of the glacial-interglacial climate signal [e.g., Weaver et al., 1998; Schneider von Deimling et al., 2006]. Support for GHG concentrations as a strong control for SST also comes from an assessment of records from tropical regions with different dynamics for simultaneous (coherent and in-phase) surface temperature changes [Medina-Elizalde and Lea, 2005]. Tropical SST records that covary with the Antarctic records of atmospheric CO₂ provide further evidence in favor of this hypothesis [Herbert et al., 2010]. Here we provide a detailed analysis which includes the influence of CO₂ on orbital-scale changes in SST at Site 1239 but limit this last analysis to the past 800 kyr, where high-resolution records of pCO₂ from ice cores exist. A coherent and in-phase relationship between atmospheric CO₂ and EEP SST would support radiative CO₂ forcing as a driver of climate in the eastern tropical Pacific.

3. Methods

3.1. Alkenone Paleothermometry

We present 342 new U₃⁷°C SST measurements from ODP Site 1239 (0°40.32’S, 82°4.86’W) to compile a high-resolution record of changes at the northern margin of the modern EEP cold tongue. New samples were taken every 8 cm from 0.07 to 27.00 meters core depth (mcd) along the suggested core splice [Shipboard Scientific Party, 2003]. Lipid ketones were extracted from ~0.5 g of dried and crushed sediment under pressurized dichloromethane in an accelerated solvent extractor, and alkenones were quantified via a Hewlett Packard gas chromatograph at the University of California, Santa Cruz (UCSC). The program used an initial oven temperature of 90°C, heating by 25°C/min until 250°C, by 1°C/min until 303°C, and by 20°C/min to 330°C, which was held for 30 min [Dekens et al., 2007]. Temperatures are estimated using the alkenone unsaturation index (C₃₇:2/(C₃₇:2 + C₃₇:3)) and empirical calibration equation of Müller et al. [1998] as cold tongue surface waters are unlikely to encounter light or nutrient limitation. Sample precision is based on replicates of a liquid consistency standard in each run and sediment standard extracted alongside each run. Long-term 1σ uncertainties of the unsaturation index are ±0.014 (±0.43°C) and ±0.009 (±0.26°C) for the sediment and liquid consistency standards, respectively. The potential for seasonal bias when using the alkenone proxy for SST has been addressed previously and shown to be of concern in a minority of regions, e.g., near the subarctic front [Rosell-Melé and Prahl, 2013] or coastal upwelling off southern Peru [Kienast et al., 2012]. In the eastern equatorial Pacific ocean there is no evidence of bias due to the seasonality of production or export of alkenones [Kienast et al., 2012] and the alkenone U₃⁷°C index in the eastern equatorial Pacific is statistically correlated with climatological mean annual SST or indeed SST in any month except February or March [Timmermann et al., 2014].

We integrated these new data with published alkenone-based SST determinations from ODP Site 1239 [Etourneau et al., 2010; Rincón-Martínez et al., 2010]. The common depth scale allowed us to create a continuous high-resolution SST record for the period of 1.5–0 Ma with a higher sampling resolution than previous records from this site. The new sample resolution for Site 1239 is on average 1.1 kyr in the period 0.9 Ma to present and 3.6 kyr in the period 1.5–0.9 Ma. Replicate (identical interval) alkenone samples from published records (analyzed in Kiel) and new records (analyzed at UCSC) show no systematic interlaboratory offset in the measured unsaturation index and are within instrumental uncertainty. We also interpolated the SST records produced in Kiel and UCSC to an even spacing and then compared the two data sets. The new data (measured at UCSC) is on average 0.17°C higher than the previously published record, which is within the 1σ uncertainty of the index.

3.2. Site 1239 Timescale

The Site 1239 age model for the period 1.5 Ma to present is based on δ¹⁸O values of benthic foraminifera Cibicidoides wuellerstorfi and Uvigerina peregrina; the C. wuellerstorfi record is adjusted by a constant offset of +0.66‰ to account for the species differences. This benthic record has an average sampling interval of 2 kyr and was completed at the College of Oceanic and Atmospheric Sciences Stable Isotope Laboratory at Oregon State University. Foraminifera picked from the >150 μm size fraction were analyzed with a
Finnigan/MAT 251 gas source mass spectrometer with an automated carbonate device. Precision for $\delta^{18}O$ is 0.05‰ and for $\delta^{13}C$ is 0.03‰. Additional published benthic isotopic measurements [Rincón-Martínez et al., 2010] were also incorporated for the period 0.5 Ma to present using the common depth scale.

The Site 1239 age model was then constructed via graphical alignment of the benthic foraminiferal $\delta^{18}O$ record with the global benthic LR04 $\delta^{18}O$ stack [Lisiecki and Raymo, 2005]. Both glacial and interglacial peaks and troughs in the two curves were matched, correcting any individual segment of the original calcareous nanoplankton occurrence age model [Shipboard Scientific Party, 2003] by <5% and the age models of Rincón-Martínez et al. [2010] and Etourneau et al. [2010] by <1% (<3 kyr). The new age model was then applied to the temperature record, from 500 to 0 ka [Rincón-Martínez et al., 2010], from 730 to 0 ka (this study), and from 1500 to 500 ka [Etourneau et al., 2010] to assemble the compiled record (Figure 2).

### 3.3. Other Climate and Climate Forcing Records

To test theories of orbital-scale tropical Pacific climate change prior to, during, and after the MPT, records that span ~1.5 Ma with a temporal resolution of <4 kyr are ideal. Only a few such high-resolution geochemical SST records exist for this region [e.g., Liu and Herbert, 2004; Medina-Elizalde and Lea, 2005]. In this study, we compare the newly compiled Site 1239 SST record with previously published tropical SSTs from Site 846 in the cold tongue, Site 806 in the western Pacific, and TR163-19 in the eastern Pacific warm pool. Wherever possible, we have limited SST comparisons to those reconstructed using the same proxy, i.e., alkenone concentrations. However, at Site 806 in the west Pacific, we make comparisons with an existing Mg/Ca-based record.
rather than the existing alkenone-based record [McClintock and Rosell-Mélè, 2005] which may be “saturated” with respect to the C\textsubscript{37:2} concentration and not fully representative of the warmer temperatures of the western warm pool. Though the SST records from Site 806 as well as a shorter record in the eastern Pacific warm pool, Site TR163-19, are derived from Mg/Ca ratios rather than alkenone concentrations, previous studies have used long-term trends in tropical Pacific SST to conclude that both of these proxies record long-term changes in mean annual SST [e.g., Dukker et al., 2008]. Here we extend this assumption but also acknowledge that a small amount of additional uncertainty is introduced when directly comparing records generated from these two different proxies. Published timescales for these additional records are likewise based on graphical alignment of benthic δ\textsuperscript{18}O values from Site 846 [Mix et al., 1995; Liu and Herbert, 2004], Site 1239 [Etourneau et al., 2010; this study], Site 806 [Bickert et al., 1993; Karas et al., 2009], and TR163-19 [Lea et al., 2000].

We compare tropical SST with climate forcing mechanisms: solar radiation, high-latitude climate, and variations in greenhouse gas concentrations. Numerical estimates of Earth’s orbital insolation are calculated from equations of motion of the solar system (Pleistocene age model uncertainty of about 0.1%) [Laskar et al., 2004]. One signal of high-latitude climate and ice volume comes from a global benthic δ\textsuperscript{18}O stack with an age model uncertainty of <2% for the majority of the record [Lisiecki and Raymo, 2005]. Another signal of high-latitude ice volume, which may be more reflective of Antarctic ice, comes from data used to derive δ\textsuperscript{18}O\textsubscript{pre} from Site 1123 east of New Zealand, which is tuned to the LR04 benthic δ\textsuperscript{18}O stack [Elderfield et al., 2012]. Carbon dioxide and methane concentrations have previously been determined for the past 800 kyr from air bubbles trapped within ice layers in Antarctic ice sheets, as compiled in Lüthi et al. [2008] and Loulergue et al. [2008], respectively, and are dated with the EDC3 timescale [Parrenin et al., 2007]. Errors in EDC3 become larger as ice layers become less distinguishable near the base of the ice sheet; however, age uncertainties for GHG records from ice cores are generally <1%, even for the oldest ice at 800 ka [Parrenin et al., 2007].

3.4. Time Series Analysis

To test for correlations between SST, insolation, and greenhouse gas concentration records, we use cross-spectral analysis to quantify coherency and phase via Crospec and Analyseries software [Palliard et al., 1996; Howell et al., 2006]. Coherency limits are set at a conservative lower bound of 80%. Rather than to directly infer causality, this approach attempts to eliminate hypotheses that are not supported by the available data. To perform cross-spectral analyses, all records were first resampled at even spacing (1 kyr resolution) using piecewise linear integration interpolation. In an initial exercise, differences between resampling methods for interpolating the time series’ (i.e., cubic spline, staircase, simple interpolation, or coarser resampling resolution) did not change the final determinations as most data were sampled at fairly even, high-resolution spacing before application of the interpolation protocol.

The phase analysis we employ incorporates uncertainties associated with the 80% confidence interval of the Blackman-Tukey method but neglects the uncertainty associated with the LR04 chronology itself as the SST and benthic δ\textsuperscript{18}O records are already explicitly tied to the LR04 record. Phase comparisons with GHG reconstructions on the EDC3 chronology, which varies from the LR04 chronology by up to 6 kyr [Parrenin et al., 2007], must have correspondingly large age uncertainties.

For each mechanism under discussion, we consider the potential forcing of the SST variations separately in the pre-MPT period (1.5–0.9 Ma, the “41 kyr world”) and the later Pleistocene (0.9–0 Ma, the “100 kyr world”). Ice age cycles in the two intervals have different characteristics, and such variability may be the result of different forcing mechanisms. The division at 0.9 Ma also splits the records into portions that are each long enough to perform robust cross-spectral analysis of Milankovitch cycles.

4. Results

The Site 1239 SST record shows clearly defined glacial-interglacial cycles (Figure 2), in which higher frequency (~41 kyr) glacial cycles appear before the MPT and lower frequency, more intense glacial cycles develop after the MPT. Such a shift in the periodicity of the Site 1239 SST record is similar to that within the nearest long alkenone-based temperature record (ODP Site 846) in the center of the cold tongue [Liu and Herbert, 2004]. Between 1.5 and 0.9 Ma, both records appear to cool by ~2°C and were consistently offset from each other by ~1.7°C. At ~900 ka, the two records begin to diverge; between 900 and 0 ka Site 846 SST maintained
Leading up to the MPT, between about 1.5 and ~0.9 Ma, EEP temperatures from ODP Sites 846 and 1239 cooled (Figure 2), which amplified the zonal tropical Pacific temperature gradient and intensified Walker circulation [de Garidel-Thoron et al., 2005; Dyez and Ravelo, 2014]. This EEP cooling is similar to that observed in the high latitudes through the early Pleistocene; that it prevails in upwelling regions may be linked to a reduction in North Atlantic Deep Water formation, a consequent reduction of associated heat transport, and/or an expansion of cool polar waters [Martínez-Garcia et al., 2010; Sexton and Barker, 2012; McClymont et al., 2013; Pena and Goldstein, 2014], which could be translated via middepth ventilation pathways to upwelling regions. Akin to other records from around the globe, Site 1239 SST includes an extended interglacial cycle starting at ~1.2 Ma, an indication that the climate system may have been approaching a threshold just before the true onset of consistent 100 kyr cycles [McClymont et al., 2013]. These similarities in cooling trends and the emergence of a “premature” 100 kyr cycle support the hypothesis that in the interval leading up to the MPT, eastern Pacific SST was linked to extratropical cooling [McClymont et al., 2008; Etourneau et al., 2010; Martínez-Garcia et al., 2010] through the thermocline [Philander and Fedorov, 2003; Liu and Herbert, 2004].

The cooling MPT climate may also have been linked to changing basal conditions beneath continental ice sheets that modified ice sheet height and/or extent and altered climate system responses to insolation forcing [e.g., Clark and Pollard, 1998]. Ice volume during glacial states seems to have increased around the MPT, through sufficient erosion of the subglacial low-friction sediment bed [Roy et al., 2004; Clark et al., 2006] and/or an increase in available precipitation [Tziperman and Gildor, 2003; McClymont et al., 2008]. More massive Northern Hemisphere ice sheets in turn could have displaced climatic zones away from the expanding glacial mass. As climatic zones shifted equatorward, the average position of the eastern Pacific warm pool and the ITCZ, while still in the Northern Hemisphere, may have also shifted southward [Chiang and Bitz, 2005]. A recent study from the west central tropical Pacific used dust tracers to interpret a more northerly position for the ITCZ from 1.1 to 0.9 Ma and a shift to a more southerly position 0.9–0.8 Ma, concurrent with increasing NH ice volume [Seo et al., 2015]. As ice sheets built up at the MPT, a southerly shift in the ITCZ could potentially have led to decreased wind-driven upwelling in the cold tongue, where Site 846 is located, and therefore could be detected by a reduction in the N-S SST gradient in the EEP (ΔSST_{1239-846}). As an analogy, the cold climate of the last glacial maximum coincided with a reduction in the N-S gradient in the EEP [e.g., Koutavas and Lynch-Stieglitz, 2005], which has been attributed to a southward shift in the

Cross-spectral analysis between the newly compiled Site 1239 SST and the Site 1239 benthic δ¹⁸O record shows that both records contain orbital periodicity and are coherent at orbital frequencies (supporting information Figure S1). However, both coherency and appropriate phasing are necessary to infer causality between forcing mechanisms; the records are divided into two portions, as different mechanisms may have dominated before and after the MPT.

5. Discussion
5.1. Mid-Pleistocene Transition

Leading up to the MPT, between about 1.5 and ~0.9 Ma, EEP temperatures from ODP Sites 846 and 1239 cooled (Figure 2), which amplified the zonal tropical Pacific temperature gradient and intensified Walker circulation [de Garidel-Thoron et al., 2005; Dyez and Ravelo, 2014]. This EEP cooling is similar to that observed in the high latitudes through the early Pleistocene; that it prevails in upwelling regions may be linked to a reduction in North Atlantic Deep Water formation, a consequent reduction of associated heat transport, and/or an expansion of cool polar waters [Martínez-Garcia et al., 2010; Sexton and Barker, 2012; McClymont et al., 2013; Pena and Goldstein, 2014], which could be translated via middepth ventilation pathways to upwelling regions. Akin to other records from around the globe, Site 1239 SST includes an extended interglacial cycle starting at ~1.2 Ma, an indication that the climate system may have been approaching a threshold just before the true onset of consistent 100 kyr cycles [McClymont et al., 2013]. These similarities in cooling trends and the emergence of a “premature” 100 kyr cycle support the hypothesis that in the interval leading up to the MPT, eastern Pacific SST was linked to extratropical cooling [McClymont et al., 2008; Etourneau et al., 2010; Martínez-Garcia et al., 2010] through the thermocline [Philander and Fedorov, 2003; Liu and Herbert, 2004].

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ITCZ due to a change in the tropical-extratropical temperature gradient [Chiang and Bitz, 2005; Broccoli et al., 2006]. However, during the long-term cooling around the MPT, the opposite is found; the secular EEP cooling from 1.5 to ~0.9 Ma culminated in a marked increase in the N-S SST difference (ΔSST1239-846) during the MPT (Figure 2). As the MPT increase in ΔSST1239-846 appears to have more to do with warming at Site 1239, rather than a change in upwelling intensity at Site 846, perhaps the mechanism causing the gradient shift has something to do with the region surrounding Site 1239.

From ~900 to 700 ka, while glacial and interglacial SSTs at Site 1239, located at the northern edge of the EEP cold tongue, warmed, SSTs at Site 846, in the center of the cold tongue, did not warm. The fact that SST at Site 846 did not increase at this time is consistent with narrower, more focused equatorial cold tongue upwelling from ~900 to 700 ka [Dyez and Ravelo, 2014]. This result has intriguing implications for understanding the factors that explain the long-term position of the ITCZ; while some ideas focus on the tropical-extratropical temperature gradient as the main factor [Chiang and Bitz, 2005; Broccoli et al., 2006], it is also thought that the shallow and cold thermocline exerts a major control on the position of the ITCZ just north of the equatorial cold tongue in the EEP today [Philander et al., 1996]. Our MPT results suggest that the cooling thermocline [Ford et al., 2012], and consequently the cooling cold tongue (as recorded at site 846), could have evolved hand in hand with a narrowly focused ITCZ just north of the equator (as indicated by increasing ΔSST1239-846). In contrast, orbital and higher frequency variations in EEP temperature records, in tropical-extratropical temperature gradients, and in the position of the ITCZ, are likely superimposed on the long-term evolution of the EEP through the MPT. In the sections below, we utilize time series analysis of tropical SST records to evaluate the evidence for climate forcing mechanisms (tropical and high-latitude insolation, ice sheet, and greenhouse gas forcing) that are proposed to have modified eastern Pacific SST on orbital timescales both before and after the MPT.

5.2. Direct Local Solar Forcing for the Cold Tongue

The idea we test first is whether changes in October equatorial insolation, which could modify interannual equatorial SST gradients by differentially heating the east and west Pacific via ITCZ and Walker circulation feedbacks [Clement et al., 1999], could also modify SST on orbital timescales. Periods of increased equatorial insolation near the autumnal equinox may have strengthened zonal equatorial SST asymmetry, reinforced Walker circulation, and led to cooler cold tongue SST over orbital timescales (thousands of years), thereby increasing the zonal SST gradient [Koutavas et al., 2002]. However, SST data from the cold tongue instead suggest warmer, rather than colder, temperatures at intervals of maximum October insolation (Figure 3), the reverse of that predicted. Although pre-MPT SST at Site 1239 is not coherent with October equatorial insolation at precession frequencies (Figure 3a), Site 846 SST, in the cold tongue, is coherent and in phase with October equatorial insolation at the 23 kyr periodicity (Figure 3b). For the period after the MPT, results show that Site 1239 SST actually leads insolation at the 23 kyr period and Site 846 is in phase with October equatorial insolation (Figures 3c and 3d). These results do not support the prediction that cold tongue SSTs and autumn insolation are anticorrelated if equatorial insolation controlled zonal SST gradients, upwelling strength, and cold tongue SSTs [e.g., Clement et al., 1999]. Rather, an alternative interpretation could be that local cold season (autumn) insolation modified cold tongue SST through direct radiative forcing.

Because the ITCZ and the equatorial front (EF) are closer to the latitude of Site 1239, this location may be more representative of temperature within the front, at the margin of the cold tongue, whereas Site 846, further south, may be more representative of the center of the cold tongue. The SST gradient between Sites 846 and 1239 can also be thought of as signaling the relative north-south position of the equatorial front, which may have been modified by local insolation. To explore the idea further, we explicitly examine variations in the SST gradient between the alkenone-based SST records from Site 1239 and Site 846 (ΔSST1239-846) as a way to monitor changes in the position of the EF, which today passes over the location of Site 1239. Thus, when the SST gradient between these sites is large, the front may be south of its present position and lie between the two sites, whereas when the gradient is small, the EF may be north of its present position, leaving both Sites 1239 and 846 in the cold tongue. The record of ΔSST1239-846 (as a proxy for the position of the EF) is not coherent or in phase with 1 October insolation before the MPT (Figure 4a). If we consider instead seasonality, the 21 March to 21 September equatorial insolation difference has the same phase as 21 March insolation.
After the MPT, the $\Delta SST_{1239-846}$ gradient is coherent with equatorial insolation at 23 kyr periodicity, but in phase only with NH summer insolation, and leads October insolation by $6 \pm 1$ kyr (Figure 4b), suggesting that the equatorial front and related ITCZ intensity may weaken with decreased NH summer insolation at the 23 kyr insolation frequency. As a further check, the SST gradient between Site TR163-19 (eastern Pacific warm pool, Figure 1) and Site 846 also weakens with decreased NH summer equatorial insolation at precession periodicity. Combined with the results above, cross-spectral analyses indicate that contrary to previous predictions from the last glacial cycle [Clement et al., 1999; Koutavas and Lynch-Stieglitz, 2003], solar heating of tropical warm pools does not seem to be responsible for increased upwelling and/or cooling in the cold tongue when we consider a larger number of glacial

Figure 3. Cross-spectral analysis (spectral density, coherence, and phase angle) of pre- and post-MPT equatorial insolation and SST at Sites 846 and 1239. Spectral density is normalized and plotted on a log scale. Dashed line represents limit of nonzero coherence at 80% confidence interval, phase only plotted when coherence is above 80% level. (a) Pre-MPT Site 1239 SST spectrum (orange) is not coherent with direct insolation (grey) at precession frequency. (b) Site 846 SST (green) is in phase with insolation (grey) at 23 kyr periodicity rather than being 180° out of phase as expected. (c) Late Pleistocene Site 1239 SST is coherent with equatorial insolation at precession frequencies, though SST leads October insolation at 23 kyr periods. (d) Similar result as pre-MPT, though both 23 and 19 kyr periodicities are in phase.
cycles. Rather, our results are consistent with increased NH summer insolation directly altering the ΔSST_{1239-846} gradient on precessional timescales, especially after the MPT (Figure 4).

In sum, while eastern Pacific tropical SST records are coherent with insolation at precession frequency, we find no evidence to support SST changing as a result of zonally asymmetric heating of the equatorial Pacific during periods of increased October insolation as suggested by numerical models [e.g., Clement et al., 1999] of the last glacial cycle, either before or after the MPT. Rather, on precessional timescales, at Sites 806, 846, and 1239 (where coherent) tropical SST and the EF are in phase with NH summer insolation especially after the MPT.

5.3. High-Latitude Orbital Insolation Forcing

Tropical SST records, especially those associated with upwelling regions, may also respond to extratropical insolation changes as mode waters, subducted at higher latitudes, upwell in the tropics [e.g., Philander and Fedorov, 2003; Liu and Herbert, 2004]. The intensity of seasonal insolation (controlled by precession) and the latitudinal distribution of annual insolation (controlled by obliquity) are two factors that could affect midlatitude oceanic conditions (such as density gradients in mode water subduction regions), and thereby impact thermocline and tropical temperatures (section 2.2). If extratropical insolation modified tropical SSTs through the thermocline, we would expect a coherent and in-phase relationship at the prevailing orbital insolation bands [e.g., Hays et al., 1976; Liu and Herbert, 2004; Lee and Poulson, 2005].

Here we use summer insolation at 65°N as an example of high-latitude insolation. In fact, mean annual insolation is coherent and in phase in the Northern and Southern Hemispheres at the obliquity frequency, so this test is not meant to narrow down a specific hemisphere but to test high-latitude insolation forcing. A number of studies suggest an important role for southern-sourced waters circulating in the tropical Pacific thermocline [Fine et al., 1987; Goodman et al., 2005; Pena et al., 2013]. Cross-spectral analysis indeed reveals links between summer solar forcing at high latitudes and tropical SSTs (section 2.2). If extratropical insolation modified tropical SSTs through the thermocline, we would expect a coherent and in-phase relationship at the prevailing orbital insolation bands [e.g., Hays et al., 1976; Liu and Herbert, 2004; Lee and Poulson, 2005].

For the interval after the MPT, in the obliquity band, summer insolation leads both Site 846 SST (phase: 3 ± 2 kyr) and Site 1239 SST by 6 ± 2 kyr (Figure 5b). These results indicate that obliquity-based solar forcing may have had less control on cold tongue SSTs in the 100 kyr world. The fact that cold tongue SST variations lag solar forcing suggests that cold tongue SSTs may be controlled by other factors in the obliquity band at this time (discussed below).

**Figure 4.** Phasing between direct equatorial insolation on 1 October (black circle at 12 o’clock position) and SST gradients (white circles) at precession bandwidth. Phase diagrams are read like a clock, e.g., records at the 9 o’clock position “lead” records at the 12 o’clock position. (a) Prior to the MPT, the equatorial front (ΔSST_{1239-846}) is not coherent with 1 October insolation (no data shown), while (b) after the MPT, the equatorial front gradients (ΔSST) lead 1 October insolation by approximately one quarter of a precession cycle. Phases (both kyr and phase angle) are reported in the supporting information.
5.4. Remote Ice Sheet Forcing

It has been suggested that the Pleistocene ice sheets may have also remotely modified tropical SSTs by shifting the average position of the ITCZ and by deflecting the latitudinal position of atmospheric bands [e.g., Deser and Wallace, 2000; Andreasen et al., 2001; Chiang and Bitz, 2005; Ruddiman, 2006; Toggweiler and Lea, 2010] on orbital timescales (section 2.3). Here we test this hypothesis by comparing a globally integrated signal of continental ice volume (LR04 stack) with EEP SST. The caveats of this choice of an indirect ice sheet signal are considered at the end of this section. If ice sheet size impacted tropical SSTs, it is expected that benthic δ¹⁸O changes and tropical SSTs should be coherent and in phase and that the relationship between benthic δ¹⁸O and tropical SST evolved when glaciations intensified around the MPT. However, only after the MPT, and only at Site 1239, are SST and benthic δ¹⁸O in phase (within error) in the obliquity band (Figure 5). This fact raises the possibility that in the obliquity band, after the MPT, SST at Site 1239 was sensitive to the position of atmospheric circulation cells and/or ice sheets, whereas SST in the heart of the cold tongue may have been influenced by extratropical insolation through changes in the thermocline as previously described. SST changes at Site 846 lead SST variability at Site 1239 by ~3–5 kyr after the MPT (Figure 5), consistent with the notion that the two localities may be influenced by different climate mechanisms at the 41 kyr periodicity. That this relationship holds after the MPT when ice sheets were thicker, though perhaps less extensive than portions of the early Pleistocene, suggests that an MPT ice volume threshold may exist for ice sheet forcing of equatorial SST at Site 1239.

At about the 100 kyr periodicity, after the MPT, phase lag relationships do not rule out a linear relationship between benthic δ¹⁸O and tropical SST at both Sites 1239 and 846 (Figure 6). This result is compatible with the theory that on longer Pleistocene timescales larger ice sheets affected global scale atmospheric patterns [Broecker and Denton, 1989; Anderson et al., 2009] which drove tropical SST changes at about the 100 kyr periodicity after the MPT, although phase lag evidence, alone, is not sufficient to prove such cause-and-effect relationships. As discussed next, pCO₂ forcing could also have played an important role in reinforcing 100 kyr tropical SST cycles.

In this section we used a benthic δ¹⁸O stack (LR04) as an indirect proxy for ice
volume, which, by incorporating many records from around the globe, may hide some of the regional complexity of ice volume, such as potential differences between the Northern and Southern Hemispheres. We also tested an alternate relative sea level record derived from benthic δ^{18}O_{sw} from Southern Hemisphere Site 1123 [Elderfield et al., 2012]. A spectral comparison between these two records is presented in supporting information Figure S2. Unfortunately, because these two records are in phase within uncertainty, we cannot distinguish with spectral analysis which signal is unique to the Southern Hemisphere. Even though differences in the two records exist (Site 1123 shows that the temperature component of benthic records may resemble a square wave and the δ^{18}O_{sw} component may provide more of the “sawtooth” character of benthic δ^{18}O calcite; Site 1123 would seem to indicate an abrupt increase in glacial ice volume at ~0.9 Ma), the two records are coherent and in phase at the obliquity periodicity throughout the past 1.5 Ma. Because the results of testing this hypothesis with either of these two records were similar, we cannot distinguish a separate southern source and instead present results using LR04.

5.5. Radiative Greenhouse Gas Forcing

Atmospheric greenhouse gas concentrations covary with Pleistocene tropical Pacific SST [Herbert et al., 2010], possibly due to radiative heating and associated tropical feedbacks [e.g., Medina-Elizalde and Lea, 2005; Dyez and Ravelo, 2013]. Further, carbon cycle changes are proposed to have acted as an internal feedback of the climate system [Ashkenazy and Tziperman, 2004; Medina-Elizalde and Lea, 2005] that may have also played a role in shifting western tropical Pacific SST periodicity at the MPT (section 2.4). Here our results conform with this theory of radiative greenhouse gas forcing having influenced eastern Pacific SST patterns.

In the time period prior to 800 ka, greenhouse gas records do not yet exist at sufficient temporal resolution for cross-spectral analysis. Nevertheless, preliminary atmospheric pCO_{2} values derived from the δ^{11}B proxy suggest a lower amplitude of variability in the period 1.5–0.8 Ma than from 0.8 Ma to present [Hönisch et al., 2009], in accord with the amplitude shift observed in benthic δ^{18}O cycles [Lisiecki and Raymo, 2005]. Higher glacial pCO_{2} concentrations prior to the MPT are in keeping with the warmer glacial intervals of that time period, though orbital-scale pCO_{2} records are still needed from before the MPT to confirm the presence, amplitude, and phasing of potential 41 kyr cycles in atmospheric pCO_{2}.

In the past 800 kyr, two out of three tropical SST records are in phase with CO_{2} concentrations (Figure 6) at the eccentricity frequency, which is consonant with the idea that radiative greenhouse gas forcing played a primary role in regulating tropical Pacific SST at eccentricity frequency after the MPT [Medina-Elizalde and Lea, 2005; Herbert et al., 2010; Dyez and Ravelo, 2013]. The lack of an eccentricity frequency signal in the power spectrum of the difference between these records is also consistent with both records being forced by the same mechanism at this frequency. These results are compatible with the idea that glacial terminations are influenced by an initial release of carbon from the deep ocean [Ruddiman, 2006; Shakun et al., 2012] and that oceanic controls for carbon fluxes into and out of the high-latitude ocean play an important role in determining the amplitude of glacial-interglacial temperature variability [Kemp et al., 2010] in the Pleistocene ice age cycles between 800 ka and present, though it is also conceivable that both CO_{2} and tropical SSTs were forced independently by another set of factors.

Atmospheric carbon dioxide records also contain components of obliquity and precession variability. At obliquity periodicity, tropical SST in both the western equatorial Pacific and EEP and benthic δ^{18}O are in phase or lead CO_{2} concentrations. In the precession bandwidth, SST records from both Sites TR163-19 and 806 lead CO_{2} (Figure 6). The fact that Site 806 SST (and Site 846 SST where coherent) tend to lead pCO_{2} at obliquity and precession indicates that GHGs may not have been a direct driver of equatorial temperature change at the 41 kyr or 23 kyr periodicities; rather, tropical climate or oceanographic changes may have influenced carbon cycle feedbacks, through possible changes in productivity, carbon uptake, or continental vegetation on these timescales.

6. Summary

We observe a shift in the frequency of SST variability in the eastern Pacific equatorial cold tongue region during the MPT, similar to other parts of the globe. The MPT may represent a response of the earth system to external forcing, which resulted in a change in the intensity and duration of glaciations [e.g., Clark and
At ~900 ka, the equatorial front at the northern margin of the cold tongue shifted as Site 1239 SST warmed over the next ~200 kyr, during and just after the MPT, even as the center of the cold tongue did not.

Spectral analysis suggests that SST variability in the tropics is not inversely related to October insolation as predicted by early climate models of the Last Glacial Maximum. Rather, at precession periodicities (19–23 kyr cycles) eastern Pacific SST is directly correlated to equatorial and extratropical radiation changes in the tropical cold season (summer–fall) rather than inversely correlated. Though the EEP cold tongue is largely influenced by the strength and temperature of upwelling or advected cool water on obliquity or eccentricity timescales, precession-scale modulations of cold tongue SST are coherent and in phase with local insolation variability both before and after the mid-Pleistocene transition.

At the 41 kyr periodicity, eastern Pacific SST variability is coherent and in phase with high-latitude insolation forcing prior to the MPT and ice sheet volume (at Site 1239, after the MPT). The pre-MPT coincidence between high-latitude insolation and changes in SST in the cold tongue supports a direct relationship between the two, possibly via modifications in midlatitude thermocline conditions which could then regulate the temperature of upwelled waters in the tropical cold tongue. SST changes at Site 1239, at the northern margin of the cold tongue, tend to lag those at Site 846, in the center of the cold tongue. Therefore, other forcing factors, which themselves lag behind solar radiation changes, such as ice volume via its effects on winds and position of the ITCZ, may have impacted SSTs at Site 1239, especially after the MPT when the northern boundary of the cold tongue could have become more narrowly focused.

At the 100 kyr periodicity, feedbacks associated with radiative greenhouse gas changes seem to be the major forcing of eastern tropical Pacific SST change. Marine and terrestrial carbon cycle interactions and the interplay of ocean overturning at high latitudes aid in the control of atmospheric CO₂ concentrations; very little power at the 100 kyr period exists in insolation records. However, high-resolution pCO₂ records are only available for the last 800 kyr; this current limitation curbs our ability to make statements about the phasing of greenhouse gas forcing for tropical SSTs in the period before the mid-Pleistocene transition though this is an active area of research.

Overall, this study is consistent with the idea that Pleistocene orbital variability in eastern tropical Pacific SST is primarily forced by insolation changes at precession periodicity, by both high-latitude insolation (Site 846) and ice sheet changes (Site 1239) at obliquity periodicity, and by radiative greenhouse gas feedbacks in the 100 kyr band.


Supplementary Text. Cross-spectral analysis between the newly compiled Site 1239 SST and the Site 1239 benthic δ¹⁸O record shows that both records have Milankovitch periodicity and are coherent at 100-, 41-, and 23-kyr frequencies (Supplementary Figure 1). To examine how the power of these frequencies changes through time, we also filtered the Site 1239 SST record for these same dominant orbital periods. Before the mid-Pleistocene transition, the amplitude of 100-kyr periodicity at Site 1239 SST is relatively low while 41-kyr and 23-kyr periodicity is relatively higher. After the mid-Pleistocene transition these trends are reversed (higher amplitude in the 100-kyr component and lower power in the 41-kyr and 23-kyr components).

Supplementary Tables. Here we present tables of phase angles for Figures 4-6. Positive values indicate SST (or minimum δ¹⁸O) lead ahead behind the insolation curve or pCO₂ concentration (negative values indicate lag), as specified in the table caption.
Table S1: Phase angles for Figure 4: Phase relative to equatorial insolation on October 1.

<table>
<thead>
<tr>
<th></th>
<th>Pre-MPT</th>
<th>0.9 Ma to present</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>23-kyr period</td>
<td>23-kyr period</td>
</tr>
<tr>
<td>Record</td>
<td>Δ SST Lead (ky)</td>
<td>Δ SST Lead (phase angle)</td>
</tr>
<tr>
<td>Δ SST&lt;sub&gt;1239-846&lt;/sub&gt;</td>
<td>6 ± 2 ky</td>
<td>93° ± 26°</td>
</tr>
<tr>
<td>Δ SST&lt;sub&gt;TR 163-846&lt;/sub&gt;</td>
<td>5 ± 2 ky</td>
<td>82° ± 24°</td>
</tr>
</tbody>
</table>

Table S2: Phase angles for Figure 5: Phase relative to 65°N Insolation on June 21. SST and benthic δ<sup>18</sup>O lag summer insolation in the northern hemisphere.

<table>
<thead>
<tr>
<th></th>
<th>Pre-MPT</th>
<th>0.9 Ma to present</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>41-kyr period</td>
<td>41-kyr period</td>
</tr>
<tr>
<td>Record</td>
<td>Lead (ky)</td>
<td>Lead (phase angle)</td>
</tr>
<tr>
<td>Site 846 SST</td>
<td>-1 ± 1 ky</td>
<td>-6° ± 11°</td>
</tr>
<tr>
<td>Site 1239 SST</td>
<td>-2 ± 2 ky</td>
<td>-21° ± 18°</td>
</tr>
<tr>
<td>Site 1090 SST</td>
<td>0 ± 1 ky</td>
<td>4 ± 11 ky</td>
</tr>
<tr>
<td>Benthic δ&lt;sup&gt;18&lt;/sup&gt;O (min. ice)</td>
<td>-6 ± 1 ky</td>
<td>-56° ± 10°</td>
</tr>
</tbody>
</table>

Table S3: Phase angles for Figure 6: Phase relative to atmospheric pCO<sub>2</sub> concentrations.

<table>
<thead>
<tr>
<th></th>
<th>100-kyr period</th>
<th>41-kyr period</th>
<th>19-23-kyr period</th>
</tr>
</thead>
<tbody>
<tr>
<td>Record</td>
<td>Lead (ky)</td>
<td>Lead (phase angle)</td>
<td>Lead (ky)</td>
</tr>
<tr>
<td>Site 806 SST</td>
<td>4 ± 5 ky</td>
<td>13° ± 15°</td>
<td>6 ± 2 ky</td>
</tr>
<tr>
<td>Site 846 SST</td>
<td>-3 ± 4 ky</td>
<td>-8° ± 13°</td>
<td>5 ± 2 ky</td>
</tr>
<tr>
<td>Site 1239 SST</td>
<td>-3 ± 2 ky</td>
<td>-12° ± 8°</td>
<td>0 ± 2 ky</td>
</tr>
<tr>
<td>TR 163-19 SST</td>
<td>2 ± 2 ky</td>
<td>19° ± 15°</td>
<td>1 ± 1 ky</td>
</tr>
<tr>
<td>Benthic δ&lt;sup&gt;18&lt;/sup&gt;O (min. ice)</td>
<td>-4 ± 1 ky</td>
<td>-13° ± 4°</td>
<td>1 ± 1 ky</td>
</tr>
</tbody>
</table>
Supplementary Figure 1. (a) Cross-spectral analysis (spectral density, coherence) of SST and benthic δ¹⁸O at Site 1239. Spectral density is normalized and plotted on a log scale. Dashed line represents limit of non-zero coherence at 80% confidence interval. (b-d) The Site 1239 SST (orange) and a benthic δ¹⁸O stack (blue) are filtered in the dominant orbital periods to assess the time-evolution of these signals.
Supplementary Figure 2. Ice sheet proxies. (a) Cross-spectral analysis (spectral density, coherence) of LR04 $\delta^{18}O$ and benthic $\delta^{18}O_{sw}$ at Site 1123 (Elderfield et al., 2012). Spectral density is normalized and plotted on a log scale. Dashed line represents limit of non-zero coherence at 80% confidence interval. (b) The original data from the two records for comparison.