Title	Enhanced upwelling in the eastern equatorial Pacific at the last five glacial terminations
Author(s)	bin Shaari, Hasrizal; Yamamoto, Masanobu; Irino, Tomohisa
Citation	Palaeogeography palaeoclimatology palaeoecology, 386, 8-15 https://doi.org/10.1016/j.palaeo.2013.03.022
Issue Date	2013-09-15
Doc URL	http://hdl.handle.net/2115/53618
Rights	Copyright © 2013 Elsevier B.V. All rights reserved.
Type	article (author version)
File Information	Shaari_Palaeo2013.pdf

[Instructions for use](#)

Enhanced upwelling in the eastern equatorial Pacific at the last five glacial terminations

Hasrizal bin Shaari*, Masanobu Yamamoto and Tomohisa Irino

Faculty of Environmental Earth Science, Hokkaido University,
Kita-10, Nishi-5, Kita-ku, Sapporo 060-0810, Japan.

*Corresponding author: Tel.:+81-11-706-2379, fax:+81-11-706-4867

E-mail address: hasrizals@ees.hokudai.ac.jp (S.Hasrizal)

Keywords: $\text{TEX}_{86}^{\text{H}}$, U_{37}^{K} , eastern equatorial Pacific, upwelling, deglaciation

Abstract

$\text{TEX}_{86}^{\text{H}}$ - and U_{37}^{K} -derived paleotemperatures, and isoprenoid glycerol dialkyl glycerol tetraether (GDGT), and alkenone concentrations were examined for ODP Site 1239 in the eastern equatorial Pacific (EEP) for the last 430 kyrs. We propose that the difference between $\text{TEX}_{86}^{\text{H}}$ - and U_{37}^{K} -derived temperatures (ΔT) and the abundance ratio of GDGTs to alkenones (GDGT/alkenone ratio) are potential upwelling indices which show a consistent results with other upwelling indices. The ΔT and GDGT/alkenone ratio were maximal during the last five deglaciations, suggesting intensified upwelling. The intensification of upwelling in the EEP coincided with those at the Peru margin and in the Southern Ocean. This coincidence suggests that the reorganization of the Southern

Hemisphere atmospheric circulation induced the intensification of the subtropical high-pressure cell, causing stronger southeast trade winds along the west coast of South America and the southern westerlies over the Southern Ocean, enhancing upwelling in both regions.

1. Introduction

The eastern equatorial Pacific (EEP) is a region between subtropical gyres of the North and South Pacific and contains the eastern terminus of the equatorial current system of the Pacific (Kessler et al., 2006). This region is important for its roles in climate variability as a result of the El Niño-Southern Oscillation (ENSO) and its significance for global carbon cycle (Fiedler and Lavin, 2006).

Glacial-interglacial changes in the oceanic condition of the EEP have been reconstructed by various studies, including of sea surface temperature (SST) (e.g., Lyle et al., 1992), salinity (e.g., Lea et al., 2000), export production (e.g., Lyle et al., 1988), and intermediate water properties (e.g., Spero and Lea, 2002; Ganeshram et al., 2000). These studies have provided evidence for an early response by the EEP to orbital forcing (e.g., Imbrie et al., 1992), and the EEP is thus thought to play an important role in amplifying climatic changes through positive feedback mechanisms.

ENSO-like variability has often been used to interpret changes in the oceanic condition of the EEP (e.g., Lea et al., 2000; Koutavas et al., 2002), but different proxy records have led to different conclusions. Some researchers, for instance, have suggested that the glacial EEP was El Niño-like based on foraminiferal Mg/Ca and $\delta^{18}\text{O}$ (e.g., Koutavas et al., 2002; Koutavas and Lynch-Stieglitz, 2003), but others have inferred a

glacial La Niña-like condition (e.g., coccolith assemblages by Beaufort et al., 2001; foraminiferal assemblages by Martinez et al., 2003; alkenones by Rincon-Martinez et al., 2010). This disagreement has been attributed to differences in the behavior of different proxies (e.g., Dubois et al., 2009).

In this paper, we present temperature records derived from $\text{TEX}_{86}^{\text{H}}$ and U_{37}^{K} for Ocean Drilling Program (ODP) Site 1239 and interpret the U_{37}^{K} and $\text{TEX}_{86}^{\text{H}}$ records for the last 430,000 years. On the basis of this interpretation, we propose the difference between $\text{TEX}_{86}^{\text{H}}$ - and U_{37}^{K} -derived temperatures and the abundance ratio of glycerol dialkyl glycerol tetraethers (GDGTs) to alkenones as potential upwelling indices and discuss the response of the EEP upwelling system to orbital forcing.

Modern physical oceanography

The zonal surface current system in the eastern tropical Pacific (ETP) consists of westward- and eastward-flowing currents (Fig. 1). The main westward currents are the North Equatorial Current (NEC; 8°N and 20°N) and the South Equatorial Current (SEC; 3°N to 10°S). The SEC originates as a combination of the waters from the North Equatorial Counter Current (NECC), the Equatorial Undercurrent (EUC), and the Peruvian Undercurrent (Kessler, 2006) through equatorial upwelling, mixing and advection. Two main lobes of the SEC are observed at latitude of about 3°S to just north of the equator. The NECC, an eastward current flows just north of the equator and is centered at about 5°N (Wyrtki, 1967; Talley et al., 2011). This current transports warmer water from the western Pacific warm pool to the ETP region. Between the SEC and the NECC there is a narrow equatorial front (EF) that separates warm low-salinity waters in

the north from cool high-salinity waters in the south (Fig. 1; Strub et al., 1998). This front is observable from July to September at about 2.5°N with a strong meridional SST gradient. In contrast, the EF position is unclear from January to April, when the southeast trades winds collapse and SST south of the Equator increases owing to reduced upwelling. The condition of the EF is correlated with the displacement of the intertropical convergence zone (ITCZ) (e.g., Pak and Zaneveld, 1974; Chelton et al., 2001; Raymond et al., 2004). The ITCZ reaches its northernmost extent in the month of August (~12°N) when southeast trade winds are stronger; the ITCZ is located closest to the equator in April (~2°N) when northeast trade winds are stronger (Waliser and Gautier, 1993).

The most influential subsurface current in this region is the EUC that flows eastward beneath the SEC. The EUC is fed by the saline New Guinea Coastal Undercurrent at its western boundary (Talley et al., 2011) and flows within the equatorial thermocline and shoal as it approaches the Galapagos Islands (Kessler, 2006). When it reaches the Galapagos Islands, it splits into two branches (Steger et al., 1998) with the main branch flowing southward to merge with the Peruvian Undercurrent, which provides a for source of the Peru coastal upwelling (Brink et al., 1983), the other branch continues to flow eastward, merging with the NECC (Wyrтки, 1967; Fieldler and Tally, 2006; Kessler, 2006).

The EEP is a region that has been impacted by coastal upwelling. Coastal upwelling in the EEP is driven by Ekman transport generated by southeast trade winds that blow along the west coast of South America (Wyrтки, 1981). The Ekman transport moves surface water offshore, away from the coastal boundary and replaces it with water from below the thermocline to maintain the mass balance. Seasonally, coastal upwelling

is at its highest intensity when the strongest southeast trade winds blow over this region in boreal summer, and is reduced when southeast trade winds are relatively weak in boreal winter (Wrytki, 1975, 1981; Kessler, 2006). The seasonal variability in the EEP is superimposed by interannual El Niño events (Wang and Fiedler, 2006), which occur every 2–7 years and last for 6–18 months (Penington et al., 2006). Hydrological conditions that characterize El Niño (La Niña) phases in the EEP are a deeper (shallower) thermocline and weaker (stronger) upwelling (Kessler, 2006).

Modern observation shows a clear seasonal and interannual SST variability in the EEP region (Fig. 2a). Seasonally, higher SST is recorded during boreal winter (February), and the lowest SST is recorded in boreal summer (August). The vertical temperature gradient is larger in boreal winter than that in boreal summer. Interannually, higher SST is observed in strong El Niño years and lower SST is observed in strong La Niña years (Fig. 2a). The thermocline depth at the study site is approximately 30–50 m (Fig. 2b).

2. Materials and methods

ODP Site 1239 (0°40.32' S, 82°4.86' W; 1414 m water depth) is located near the eastern crest of the Carnegie Ridge and ~120 km off the coast of Ecuador (Fig. 1). The sediments are dominated by light to dark olive gray foraminifera-nannofossil ooze with varying amounts of diatoms, clay, and micrite (Mix et al., 2003). The age-depth model of this core was established by Rincon-Martinez et al. (2010) based on correlation of the $\delta^{18}\text{O}$ record of the benthic foraminifera *Cibicidoides wuellerstorfi* with the LR04 global stack (Lisiecki and Raymo, 2005). In total, 236 samples were taken from 0.02 to 14.73 meters composite depth (mcd) at 2–10 cm intervals.

Extraction and separation of lipids followed the modified method of Yamamoto et al. (2000) and Yamamoto et al. (2008). Freeze-dried and homogenized samples (~2 g) were extracted using an Accelerated Solvent Extractor 200 (ASE 200, DIONEX) with a mixture of dichloromethane and methanol (6/4 v/v) at 100°C. The extract was separated into four fractions of lipid sequences in order of polarity, F1 (3 ml hexane), F2 (3 ml 3/1 v/v hexane–toluene), F3 (4 ml toluene), and F4 (3 ml 3/1 v/v toluene–methanol), by column chromatography (SiO₂ with 5% distilled water; i.d., 5.5 mm; length, 45 mm).

The F3 fraction was analyzed with a Hewlett–Packard Model 6890 gas chromatograph with on-column injection and electronic pressure control inlet systems and a flame ionization detector (FID). Helium was used as carrier gas with the flow velocity maintained at 30 cm.s⁻¹. The column was a Chrompack CP-Sil5CB capillary (60 m; i.d., 0.25 mm; thickness, 0.25 µm). The oven temperature was programmed from 70 to 290°C at 20°C min⁻¹, from 290 to 310°C at 0.5°C min⁻¹, and then held for 30 min. Quantification of di- and tri-unsaturated C₃₇ alkenones were achieved by comparing the peak areas with that of an internal standard (*n*-C₃₆H₇₄) on the gas chromatogram.

The alkenone unsaturation index U^K_{37'} was computed from the concentrations of di-unsaturated (C_{37:2}MK) and tri-unsaturated (C_{37:3}MK) alkenones using the following equation by Prahl et al. (1988):

$$U_{37'}^K = [C_{37:2}MK] / ([C_{37:3}MK] + [C_{37:2}MK])$$

The temperature was calculated according to an equation derived by Prahl et al. (1988) based on experimental results for cultured strain 55a of *Emiliania huxleyi*:

$$U_{37}^{K'} = 0.034T + 0.039$$

where T = temperature (°C). Analytical accuracy was 0.24°C in our laboratory.

An aliquot of F4 was dissolved in hexane-2-propanol (99/1, v/v). GDGTs were analyzed using high-performance liquid chromatography–mass spectrometry (HPLC-MS) with an Agilent 1100 HPLC system connected to a Bruker Daltonics micrOTOF-HS time-of-flight mass spectrometer. Separation was conducted using a Prevail Cyano column (2.1 x 150 mm, 3µm; Alltech) maintained at 30°C following the method of Hopmans et al. (2000) and Schouten et al. (2007). Detection was achieved by atmospheric pressure positive ion chemical ionization–mass spectrometry (APCI-MS) with full scan mode (m/z 500–1500). Compounds were identified by comparing mass spectra and retention times with those of GDGT standards (formed from the main phospholipids of *Thermoplasma acidophilum* via acid hydrolysis).

Quantification was achieved by integrating the summed-peak areas in the (M+H)⁺ and the isotopic (M+H+1)⁺ ion traces and comparing these to the peak area of an internal standard (C₄₆ GDGT) in the (M+H)⁺ ion trace, following to the method of Huguet et al. (2006). The correction value of ionization efficiency between GDGTs and the internal standard was obtained by comparing the peak areas of *T. acidophilum*-derived mixed GDGTs and C₄₆ GDGT of known amounts. The standard deviation of a replicate analysis was 3.0% of the concentration for each compound. Concentration TEX₈₆^H (applicable in warm water) were calculated from the concentrations of GDGT-1, GDGT-2,

GDGT-3, and a regioisomer of crenarchaeol using the following expressions (Schouten et al., 2002; Kim et al., 2010):

$$\text{TEX}_{86} = ([\text{GDGT-2}] + [\text{GDGT-3}] + [\text{Crenarchaeol regioisomer}]) / ([\text{GDGT-1}] + [\text{GDGT-2}] + [\text{GDGT-3}] + [\text{Crenarchaeol regioisomer}])$$

$$\text{TEX}_{86}^{\text{H}} = \log (\text{TEX}_{86})$$

Temperature was calculated according to the following equation based on a global core-top calibration (Kim et al., 2010):

$$T = 68.4\text{TEX}_{86}^{\text{H}} + 38.6 \text{ (when } T > 15^{\circ}\text{C)}$$

where T = temperature (°C). The analytical accuracy was 0.45 °C in our laboratory.

3. Results

3.1 GDGTs and TEX_{86}

The isoprenoid GDGTs detected in ODP 1239 sediments consist of caldrarchaeol (GDGT-0), GDGT-1, GDGT-2, GDGT-3, crenarchaeol, and its regioisomer (Appendix I). The total concentration of isoprenoid GDGTs in sediment varied between 0.6 and 12.8 $\mu\text{g.g}^{-1}$ with an average of 5.81 $\mu\text{g.g}^{-1}$ (Fig. 3b). The relative abundances of different isoprenoid GDGTs were nearly uniform with range of 37–54% for crenarchaeol, 26–35% for caldrarchaeol and 15–35% for others.

The TEX₈₆^H-derived temperature of the core-top sample (25.1°C) agreed with the mean annual SST at the study site (24.5°C, Locarnini et al., 2010). TEX₈₆^H-derived SST varied between 20.2 and 27.2°C and was generally higher during interglacials and lower during glacials (Fig. 3a).

The branched isoprenoid tetraether (BIT) index, an indicator of soil bacteria contribution (see Hopmans et al., 2004), varied between 0.01 and 0.06 (Fig. 3d) suggesting a low contribution of soil organic matter in the study samples. Weijers et al. (2006) noted that samples having high BIT (>0.4) may show anomalously high TEX₈₆^H-derived temperatures, but this concern was not relevant for the samples used in this study.

3.2. Alkenones and U_{37}^K

The total concentration of C₃₇-C₃₉ alkenones in sediment varied between 0.5 and 28.7 µg.g⁻¹ with an average of 8.9 µg.g⁻¹ (Fig. 3b). The alkenone concentration tended to be higher in the interval between 400 ka and 240 ka than in the intervals between 430 and 400 ka and between 240 and 0 ka.

The U_{37}^K -derived temperature of the core-top sample (25.6°C) agreed with the mean annual SST. U_{37}^K -derived SST varied between 21.5 and 26.6°C and was generally higher in interglacials and lower in glacials (Fig. 3a). The U_{37}^K record obtained in this study was nearly identical to a record for the study site by Rincon-Martinez et al. (2010).

4. Discussion

4.1. Difference in proxy-derived temperatures

The variation of $\text{TEX}_{86}^{\text{H}}$ -derived temperature is roughly consistent with those of the U_{37}^{K} '-derived temperature at the study site (Fig. 3a), but significant difference was observed in the intervals of late MIS 11, and MIS 10, and MIS-6 when the U_{37}^{K} '-derived temperature was a maximum of 5.5°C higher than the $\text{TEX}_{86}^{\text{H}}$ -derived temperature (Fig. 3a).

Dubois et al. (2009) and Kienast et al. (2012) assumed that U_{37}^{K} ' reflects mean annual SST because the U_{37}^{K} '-derived temperature in EEP core-top sediments corresponded to mean annual SST. A sediment trap study at two sites in the central tropical Pacific showed no significant difference in the sinking flux of alkenone producers (*Emiliana huxleyi* and *Gephyrocapsa oceanica*) between strong and weak El Niño periods (Broerse, 2000), suggesting that the production of alkenone is not sensitive to upwelling intensity. We thus assume that U_{37}^{K} ' does reflect the mean annual SST at the study site.

The behavior of Thaumarchaeota and the production of GDGTs are not fully clear in the EEP. Thaumarchaeota (GDGTs producer) are ubiquitous and abundant throughout the seawater column (e.g., Massana et al., 2000; Karner et al., 2001). In the central equatorial Pacific, GDGTs are mainly produced in the thermocline layer (TL) (Turich et al., 2007). Recent case studies assumed that the $\text{TEX}_{86}^{\text{H}}$ -derived temperatures in EEP sediments reflect the temperature of the thermocline (30–50 m) rather than SSTs (Ho et al., 2011; Seki et al., 2012). Thaumarchaeota in marine environments have been recognized to be both heterotrophs (e.g., Ouverney and Fuhrman, 2000; Agogue et al., 2008; Zhang et al., 2009) and chemoautotrophic nitrifiers (e.g., Könneke et al., 2005; Hallam et al., 2006). Organic matter and NH_3 are produced by phytoplankton and by the

decay of organic matter in surface and subsurface water, which explains why Thaumarchaeota are produced in both the surface mixed layer (SML) and TL. We thus assume that $\text{TEX}_{86}^{\text{H}}$ reflects a mixed temperature signal from the SML and TL (Fig.4). The production of Thaumarchaeota is fueled by the supply of organic matter and NH_3 . Both are more enhanced by phytoplankton production in upwelling periods. Yamamoto et al. (2012) observed that enhanced sinking flux of GDGTs is linked with phytoplankton bloom in the mid-latitude northwestern Pacific. GDGT abundance thus may reflect primary production and upwelling intensity.

$\text{TEX}_{86}^{\text{H}}$ showed higher temperatures than U_{37}^{K} during the some deglaciations (Fig. 3a), but this does not mean that the integrated SST of the SML and TL was higher than the SST of the SML. The calibration of $\text{TEX}_{86}^{\text{H}}$ to SST was conducted by comparing core-top $\text{TEX}_{86}^{\text{H}}$ with mean annual SST (Kim et al., 2010). If the phenomenon of $\text{TEX}_{86}^{\text{H}}$ recording both the SST and thermocline temperatures is common in tropical oceans, then calibration requires a comparison between core top $\text{TEX}_{86}^{\text{H}}$ and integrated temperatures of the SML and TL; this calibration should give cooler estimates. The temperature reversal during the last deglaciation is thus attributed to the overestimation of $\text{TEX}_{86}^{\text{H}}$ -derived temperature.

4.2. GDGT/alkenone ratio and ΔT as upwelling indices

The relative abundance of isoprenoid GDGTs to alkenones (GDGT/alkenone ratio) was enhanced during the last five deglaciations (Fig. 5a), suggesting an enhanced production of GDGTs. When upwelling intensifies, GDGT production increases due to increasing NH_3 and organic matter. In contrast, when upwelling weakens, GDGT

production decreases. The GDGT/alkenone ratio can thus be used as an index of upwelling intensity.

The difference between $\text{TEX}_{86}^{\text{H}}$ - and U_{37}^{K} -derived temperatures (ΔT) was computed by subtracting U_{37}^{K} -derived SST from $\text{TEX}_{86}^{\text{H}}$ -derived temperature ($\Delta T = \text{TEX}_{86}^{\text{H}} - \text{U}_{37}^{\text{K}}$). U_{37}^{K} reflects the temperature of the SML and $\text{TEX}_{86}^{\text{H}}$ reflects integrated temperatures from the SML and the TL. When upwelling intensifies, the temperature gradient between the SML and TL decreases (Fig. 4), and ΔT shifts in a positive direction. In contrast, when upwelling weakens, the temperature gradient between the SML and TL increases, and ΔT shifts in a negative direction. We thus assume that ΔT is a potential index of upwelling intensity.

ΔT varied between -6.2 and 4.1°C and showed maxima at 15, 50, 127, 213, 243, 260, 274, 310, 330, 340 and 427 ka. The maxima at 15, 127, 243, 340, and 427 ka correspond to glacial terminations (Fig. 5a). Minimal peaks of ΔT occurred at 33, 86, 179, 237, 289, and 386 ka. The variation in ΔT is very similar to that in the GDGT/alkenone ratio, although there are some mismatches in MIS 8 and MIS 11. This correspondence suggests that both are robust indices of upwelling intensity.

Positive ΔT and an elevated GDGT/alkenone ratio at the study site during deglaciations are associated with heavier $\delta^{18}\text{O}$ of subsurface-dwelling foraminifera (Pena et al., 2008) and increased export production (Pedersen, 1983; Lyle et al., 1988; Kienast et al., 2006) in the EEP. Pena et al. (2008) showed that thermocline water $\delta^{18}\text{O}$ ($\text{DT-}\delta^{18}\text{O}_{\text{sw}}$) at ODP Site 1240, reconstructed from the subsurface-dwelling foraminifera *Neogloboquadrina dutertrei*, was maximized during the last three deglaciations (Fig. 5d). This suggests intensified upwelling in those periods. Abrupt increases in organic carbon

content during deglaciations were reported from sites P6 (Pedersen, 1983), V19-28 (Lyle et al., 1988), and ME0005A-24JC and 27JC (Kienast et al., 2006) in the EEP (Fig. 5b), suggesting that export production was maximized during the last two deglaciations. The elevated ΔT and GDGT/alkenone ratio at the study site indicate not only the intensification of local upwelling, but also the intensification of regional upwelling associated with thermocline shoaling and enhanced export production in the EEP during deglaciations.

4.3. Hydrological evolution in the EEP

The ΔT record mirrors sedimentary $\delta^{15}\text{N}$ records from the Peru margin (Fig. 5c), which have been suggested to reflect the intensity of denitrification regulated by Peruvian coastal upwelling (Ganeshram et al., 2000). The trend in $\delta^{15}\text{N}$ at the Peru margin was slightly different from those in the EEP (Dubois and Kienast, 2011) and at the Mexican margin (Ganeshram et al., 2000) (Fig. 5c). The maxima of $\delta^{15}\text{N}$ at terminations are significant at the Peru margin but not in the EEP or at the Mexican margin, suggesting that $\delta^{15}\text{N}$ in the eastern Pacific margin was determined by the denitrification in the Peru margin and modified by local factors (Robinson et al., 2009). The correspondence between ΔT and the Peru margin $\delta^{15}\text{N}$ records suggests that the upwelling at the study site was closely linked with Peruvian coastal upwelling. The study site is located in a region influenced by the coastal upwelling system (Wrytki, 1981; Pennington et al., 2006; Talley et al., 2011). Because the southeast trade winds are a principal agent driving coastal upwelling along the west coast of South American continent (Wrytki, 1981;

Kessler, 2006), it is highly likely that the southeast trade winds intensified during deglaciations owing to the stronger South Pacific High.

The paleo-position of the ITCZ was approximated using dust fluxes across the equator over the last 30 ka (McGee et al., 2007). The results of that analysis suggest that the ITCZ did not shift southward during the last deglaciation. Xie and Marcantonio (2012) precisely estimated the paleo-position of the ITCZ using neodymium isotopes (ϵ_{Nd}) derived from transect dust obtained by McGee et al. (2007). The average ϵ_{Nd} values from the last glacial and Holocene show similar gradients throughout the equatorial transect, but the latitudinal gradient was stronger, and a steeper interval was evident during the last deglaciation between 5°N and 7°N. This suggests more northerly mean position of the ITCZ.

Yamamoto et al. (2007) reconstructed the intensity of the California Current during the last 150,000 years and showed that the subtropical high-pressure cell in the North Pacific weakened during the last two deglaciations. Lyle et al. (2012) suggested that high precipitation in the Great Basin of the western United States during the last deglaciation was not caused by the southward shift of westerly storms, but instead by the northward transport of moist air masses from the tropical Pacific because of the weaker North Pacific High. This presumes that the northeast trade winds were not intensified under the condition of the weaker North Pacific High.

The stronger South Pacific High, combined with the weaker North Pacific High and northward shift of the ITCZ during the last deglaciation was an asymmetrical atmospheric phenomenon between the Northern and Southern hemispheres. This anti-phase variation in the subtropical high-pressure cells of both hemispheres was

presumably caused by changes in the heat balance between the hemispheres (Fig. 6).

The ENSO model has been applied to understand hydrological evolution of the EEP (e.g., Lea et al., 2000; Koutavas et al., 2002; Koutavas and Lynch-Stieglitz, 2003; Martinez et al., 2003; Pena et al., 2008; Rincon-Martinez et al., 2010). Pena et al. (2008) proposed the deep thermocline seawater $\delta^{18}\text{O}$ (DT- $\delta^{18}\text{O}_{\text{sw}}$) based on *Neogloboquadrina dutertrei* $\delta^{18}\text{O}$ at Site 1240 and suggested that EEP hydrology was characterized by a La Niña-like condition during deglaciations. However, the zonal gradient of SST was inconsistent with a La Niña-like state during the last deglaciation (Fig. 5e). The DT- $\delta^{18}\text{O}_{\text{sw}}$ at ODP Site 1240 showed maximum peaks during deglaciations (Pena et al., 2008), but the Mg/Ca-SST, between the western and eastern Pacific did not show a large temperature gradient typical of La Niña (Lea et al., 2000). Also, the weaker North Pacific High evidenced during the last deglaciation (Yamamoto et al., 2007; Lyle et al., 2012) is not consistent with a La Niña-like state; a weaker North Pacific High is typical of the modern El Niño condition (Bogad and Lynn, 2001). We thus suggest that intensified upwelling shown by enhanced DT- $\delta^{18}\text{O}_{\text{sw}}$ at Site 1240 was not linked to a La Niña-like state, and an ENSO analogy cannot to be applied to explain hydrological conditions in the Pacific during the last deglaciation.

The intensification of upwelling in the EEP and the Peru margin during the last deglaciation coincided with intensification of upwelling in the Southern Ocean (Toggweiler et al., 2006; Anderson et al., 2009). Because upwelling in the Southern Ocean is regulated by the position of the southern westerlies (Russell et al., 2006; Toggweiler et al., 2006), the synchronous intensification of upwelling systems in the EEP, the Peru margin, and the Southern Ocean suggests that the reorganization of atmospheric

circulation in the Southern Hemisphere induced the intensification of the subtropical high-pressure cell, causing stronger southeast trade winds along the west coast of South America and southern westerlies over the Southern Ocean, enhancing upwelling in both regions.

The intensification of the South Pacific High caused southern westerlies to move poleward and the ITCZ to shift northward during deglaciations (Fig. 6). In response, the center of upwelling moved northward and cold tongue upwelling in the EEP area intensified. The stronger South Pacific High during the last deglaciation caused a drier climate in the Patagonia region of South America (de Porras et al., 2012), and the weaker North Pacific High caused a wetter climate in the Great Basin of the western United States (Lyle et al., 2012). This perspective is useful for understanding the hydrological and climatological evolution of the eastern Pacific region.

5. Conclusions

The abundance ratio of GDGTs to alkenone (GDGT/alkenone ratio) and difference between $\text{TEX}_{86}^{\text{H}}$ - and U_{37}^{K} -derived temperature (ΔT) can be used as upwelling indices in the EEP. Our new data show that intensification of upwelling occurred in the EEP at each of the last five glacial terminations. The result suggests that the intensification of upwelling was a common phenomenon in the EEP at glacial terminations. The similar timing of intensified upwelling in the EEP, the Peru margin, and the Southern Ocean suggests an intensification of the South Pacific High during deglaciations. This new perspective can help explain the hydrological evolution of the eastern Pacific region during deglaciations.

366

367 **Acknowledgements**

368 The Ocean Drilling Program Leg 202 provided the samples used in this study. We
369 thank Daniel Rincon Martinez for providing us with data on the age and foraminiferal
370 oxygen isotopes of the study core. We also thank Patrizia Ziveri, Alexandra Broerse, and
371 Kyoko Hagino for providing information on coccolithophores in the EEP, Leopoldo Pena
372 for data regarding the thermocline water $\delta^{18}\text{O}$ from ODP Site 1240, Eric Galbraith for
373 data of $\delta^{15}\text{N}$ records from core NH22P and CD38-02, and Tatsufumi Okino, Keiko
374 Ohnishi and Tadamichi Oba for their analytical assistance.

375

376 **References**

- 377 Agogu, H., Brink, M., Dinasquet, J., Herndl, G.J., 2008. Major gradients in putatively
378 nitrifying and non-nitrifying Archaea in the deep North Atlantic. *Nature* 456, 788–
379 792.
- 380 Anderson, R.F., Ali, S., Bradtmiller, L.I., Nielsen, S.H.H., Fleisher, M.Q., Anderson,
381 B.E., Burckle, L.H., 2009. Wind-driven upwelling in the Southern Ocean and the
382 deglacial rise in atmospheric CO_2 . *Science* 323, 1443–1448.
- 383 Beaufort, L., de Garidel-Thoron, T., Mix, A.C., Pisias, N.G., 2001. ENSO-like forcing on
384 oceanic primary production during the late Pleistocene. *Science* 293, 2440–2444.
- 385 Bograd, S., Lynn, R.J., 2001. Physical-biological coupling in the California Current
386 during the 1997–99 El Nio-La Nia cycle. *Geophysical Research Letters* 28, 275–
387 278.
- 388 Brink, K.H., Halpern, D., Huyer, A., Smith, R.L., 1983. The physical environment of the

389 Peruvian upwelling system. *Progress in Oceanography* 12, 285–305.

390 Broerse, A.T.C., 2000. Coccolithophore export production in the selected ocean
391 environments, Ph.D. thesis, Faculty of Earth Science, Vrije Universiteit, Amsterdam,
392 Netherlands.

393 Chelton, D.B., Esbensen, S.K., Schlax, M.G., Thum, N., Freilich, M.H., Wentz, F.J.,
394 Gentemann, C.L., McPhaden, M.J., Schopf, P.S., 2001. Observations of coupling
395 between surface wind stress and sea surface temperature in the eastern tropical
396 Pacific. *Journal of Climate* 14, 1479–1498.

397 de Porras, M.E., Maldonado, A., Abarzúa, A.M., Cárdenas, M.L., Francois, J.P., Martel-
398 Cea, A., Stern, C.R., Méndez, C., Reyes, O., 2012. Postglacial vegetation, fire and
399 climate dynamics at Central Chilean Patagonia (Lake Shaman, 44°S). *Quaternary*
400 *Science Reviews* 50, 71–85.

401 Dubois, N., Kienast, M., Normandeau, C., Herbert, T.D., 2009. Eastern equatorial
402 Pacific cold tongue during the last glacial maximum as seen from alkenone
403 paleothermometry. *Paleoceanography* 24, PA4207.

404 Dubois, N., Kienast, M., 2011. Spatial reorganization in the equatorial divergence in the
405 eastern tropical Pacific during the last 150 kyr. *Geophysical Research Letters* 38,
406 L16606.

407 Fiedler, P.C., Lavin, M.F., 2006. Introduction: a review of eastern tropical Pacific
408 oceanography. *Progress in Oceanography* 69, 94–100.

409 Fiedler, P.C., Talley, L.D., 2006. Hydrography of the eastern tropical Pacific: a review.
410 *Progress in Oceanography* 69 (2–4), 143–180.

411 Ganeshram, R.S., Pedersen, T.F., Calvert, S.E., McNeil, G.W., Fontugne, M.R., 2000.
 412 Glacial-interglacial variability in denitrification in the world's oceans: causes and
 413 consequences. *Paleoceanography* 15, 361–376.

414 Hallam, S.J., Mincer, T.J., Schleper, C., Preston, C.M., Roberts, K., Richardson, P.M.,
 415 DeLong, E.F., 2006. Pathway of carbon assimilation and ammonia oxidation
 416 suggested by environmental genomic analyses of marine Crenarchaeota. *PLoS*
 417 *Biology* 4, 520–536.

418 Ho, S. L., Yamamoto, M., Mollenhauer, G., Minagawa M., 2011. Core top TEX₈₆ values
 419 in the south and equatorial Pacific. *Organic Geochemistry* 42, 94–99.

420 Hopmans, E.C., Schouten, S., Pancost, R., van der Meer, M.T.J., Sinninghe Damsté, J.S.,
 421 2000. Analysis of intact tetraether lipids in archaeal cell material and sediments by
 422 high performance liquid chromatography/atmospheric pressure chemical ionization
 423 mass spectrometry. *Rapid Communications in Mass Spectrometry* 14, 585–589.

424 Hopmans E.C., Weijers, J.W.H., Schefuß, E., Herfort, L., Sinninghe Damsté, J.S.,
 425 Schouten, S., 2004. A novel proxy for terrestrial organic matter in sediments based
 426 on branched and isoprenoid tetraether lipids. *Earth and Planetary Science Letters* 224,
 427 107–116.

428 Huguet, C., Hopmans, E.C., Febo-Ayala, W., Thompson, D.H., Sinninghe Damsté, J.S.,
 429 Schouten, S., 2006. An improved method to determine the absolute abundance of
 430 glycerol dibiphytanyl glycerol tetraether lipids. *Organic Geochemistry* 37, 1036–
 431 1041.

432 Imbrie, J., Boyle, E.A., Clemens, S.C., Duffy, A., Howard, W.R., Kukla, G., Kutzbach, J.,
 433 Martinson, D.G., McIntyre, A., Mix, A.C., Molfino, B., Morley, J.J., Peterson, L.C.,

434 Piasias, N.G., Prell, W.L., Raymo, M.E., Shackleton, N.J., Toggweiler, J.R., 1992. On
 435 the structure and origin of major glaciation cycles. I. Linear responses to
 436 Milankovitch forcing. *Paleoceanography* 7, 701–738.

437 Karner, M.B., DeLong, E.F., Karl, D.M., 2001. Archaeal dominance in the esopelagic
 438 zone of the Pacific Ocean. *Nature* 409, 507–510.

439 Kessler, W.S., 2006. The circulation of the eastern tropical Pacific: a review. *Progress in*
 440 *Oceanography* 69, 181–217.

441 Kienast, M., Kienast, S., Calvert, S., Eglinton, T., Mollenhauer, G., Francois, R., Mix, A.,
 442 2006. Eastern Pacific cooling and Atlantic overturning circulation during the last
 443 deglaciation. *Nature* 443, 846–849.

444 Kienast, M., MacIntyre, G., Dubois, N., Higginson, S., Normandeau, C., Chazen C.R.,
 445 2012. Alkenone unsaturation in surface sediments from the eastern equatorial
 446 Pacific: Implications for SST reconstructions. *Paleoceanography* 27, PA1210.

447 Kim, J.H., van der Meer, J., Schouten, S., Helmke, P., Willmott, V., Sangiorgi, F., Koc,
 448 N., Hopmans, E.C., Sinninghe Damsté, J.S., 2010. New indices and calibrations
 449 derived from the distribution of crenarchaeol isoprenoid tetraether lipids:
 450 implications for past sea surface temperature reconstructions. *Geochimica et*
 451 *Cosmochimica Acta* 74, 4639–4654.

452 Könneke, M., Bernhard, A.E., de la Torre, J.R., Walker, C.B., Waterbury, J.B., Stahl,
 453 D.A., 2005. Isolation of an autotrophic ammonia-oxidizing archaeon. *Nature* 437,
 454 543–546.

455 Koutavas, A., Lynch-Stieglitz, J., 2003. Glacial-interglacial dynamics of the eastern
 456 equatorial Pacific cold tongue. Intertropical Convergence Zone system reconstructed
 457 from oxygen isotope records. *Paleoceanography* 18, 1089.

458 Koutavas, A., Lynch-Stieglitz, J., Marchitto, T.M., Sachs, J.P., 2002. El Niño-like pattern
 459 in ice age tropical Pacific sea surface temperature. *Science* 297, 226–230.

460 Lea, D.W., Pak, D.K., Spero, H.J., 2000. Climate impact of late quaternary equatorial
 461 Pacific sea surface temperature variations. *Science* 289, 1719–1724.

462 Lisiecki, L.E., Raymo, M.E., 2005. A Pliocene-Pleistocene stack of 57 globally
 463 distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20, PA1003.

464 Locarnini, R.A., Mishonov, A.V., Antonov, J.I., Boyer, T.P., Garcia, H.E., Baranova, O.
 465 K., Zweng, M.M., Johnson, D.R., 2010. In: Levitus, S. (Ed.), *World Ocean Atlas*
 466 2009, 1, Temperature, NOAA Atlas NESDIS, 68, U.S. Government Printing Office,
 467 Washington, D.C., pp. 184.

468 Lyle, M., Murray, D.W., Finney, B.P., Dymond, J., Pobbins, J.M., Brooksforce, K., 1988.
 469 The record of Late Pleistocene biogenic sedimentation in the eastern tropical Pacific
 470 Ocean. *Paleoceanography* 3, 39–59.

471 Lyle, M., Prahl, F.G., Sparrow M.A., 1992. Upwelling and productivity changes inferred
 472 from a temperature record in the central equatorial Pacific. *Nature* 355, 812–815.

473 Lyle, M., Heusser, L., Ravelo, C., Yamamoto, M., Barron, J., Diffenbaugh, N.S., Herbert,
 474 T., Andreasen, D., 2012. Out of the Tropics: The Pacific, Great Basin Lakes, and
 475 Late Pleistocene water cycle in the western United States. *Science* 337, 1629–1633.

476 Martinez, I., Keigwin, L., Barrows, T.T., Yokoyama, Y., Southon, J., 2003. La Niña-like
 477 conditions in the eastern equatorial Pacific and a stronger Choco jet in the northern
 478 Andes during the last glaciations. *Paleoceanography* 18, 1033.

479 Massana, R., Delong, E. F., Pedros-Alio, C., Murray, A. E. Preston, C. M. ,2000. A few
 480 cosmopolitan phylotypes dominate planktonic archaeal assemblages in widely
 481 different oceanic provinces. *Applied of Environmental Microbiology* 66, 1777–1787.

482 Medina-Elizalde, M., Lea, D.W., 2005. The Mid-Pleistocene transition in the tropical
 483 Pacific. *Science* 310, 1009–1012.

484 McGee, D., Marcantonio F., Lynch-Stieglitz, J., 2007. Deglacial changes in dust flux into
 485 the eastern equatorial Pacific. *Earth and Planetary Science Letters* 257, 215–230.

486 Mix, A.C., Tiedemann, R., Blum, P., et al., 2003. Proceeding of the Ocean Drilling
 487 Program, Initial Reports, 202: College Station, TX (Ocean Drilling Program), 1–93.

488 Ouverney, C.C., Fuhrman, J.A., 2000. Marine planktonic archaea take up amino acids.
 489 *Applied Environmental Microbiology* 66, 4829–4833.

490 Paillard, D., Labeyrie, L., Yiou, P., 1996. Macintosh program performs time-series
 491 analysis. *Eos Transaction AGU* 77, 379.

492 Pak, H., Zaneveld, J.R.V., 1974. Equatorial front in the eastern Pacific Ocean. *Journal of*
 493 *Physical Oceanography* 4, 570–578.

494 Pedersen, T.F., 1983. Increased productivity in the eastern equatorial Pacific during the
 495 last glacial maximum (19,000 to 14,000 yr. B.P.). *Geology* 11, 16–19.

496 Pena, L.D., Cacho, I., Ferretti, P., Hall, M.A., 2008. El Niño-Southern Oscillation-like
 497 variability during glacial terminations and interlatitudinal tele-connections.
 498 *Paleoceanography* 23, PA3101.

499 Pennington, J.T., Mahoney, K.L., Kuwahara, V.S., Kolber, D.D., Calienes, R., Chavez,
 500 F.P., 2006. Primary production in the eastern tropical Pacific: a review. *Progress in*
 501 *Oceanography* 69 (2 – 4), 285–317.

502 Prahl, F.G., Muehlhausen, L.A., Zahnke, D.L., 1988. Further evaluation of long chain
 503 alkenones as indicators of paleoceanographic conditions. *Geochimica et*
 504 *Cosmochimica Acta* 52, 2303–2310.

505 Raymond, D.J., Esbensen, S.K., Paulson, C., Gregg, M., Bretherton, C.S., Petersen, W.A.,
 506 Cifelli, R., Shay, L.K., Ohlmann, C., Zudeima, P., 2004. EPIC2001 and the coupled
 507 ocean-atmosphere system of the tropical east Pacific. *Bulletin of American*
 508 *Meteorological Society* 85, 1341–1354.

509 Reynolds, R.W., Rayner, N.A., Smith, T.M., Stokes, D.C., Wang, W., 2002. An
 510 improved in situ and satellite SST analysis for climate. *Journal of Climate* 15, 1609–
 511 1625.
 512 <[http://iridl.ldeo.columbia.edu/SOURCES/.NOAA/.NCEP/.EMC/.CMB/.GLOBAL/.](http://iridl.ldeo.columbia.edu/SOURCES/.NOAA/.NCEP/.EMC/.CMB/.GLOBAL/.Reyn_SmithOIv2/.monthly/)
 513 [Reyn_SmithOIv2/.monthly/](http://iridl.ldeo.columbia.edu/SOURCES/.NOAA/.NCEP/.EMC/.CMB/.GLOBAL/.Reyn_SmithOIv2/.monthly/)>

514 Rincon-Martínez, D., Lamy, F., Contreras, S., Leduc, G., Bard, E., Saukel, C., Blaz, T.,
 515 Mackensen, A., Tiedeman, R., 2010. More humid interglacials in Ecuador during the
 516 past 500 kyr linked to latitudinal shifts of the equatorial front and Intertropical
 517 Convergence Zone in the eastern equatorial Pacific. *Paleoceanography* 25, PA2210.

518 Robinson, R.S., Martinez, P., Pena, L.D., Cacho, I., 2009. Nitrogen isotopic evidence for
 519 deglacial changes in nutrient supply in the eastern equatorial Pacific.
 520 *Paleoceanography* 24, PA4213.

521 Russell, J.L., Dixon, K.W., Gnanadesikan, A., Stouffer, R.J., Toggweiler, J.R., 2006. The
 522 Southern hemisphere westerlies in a warming world: propping open the door to the
 523 deep ocean. *Journal of Climate* 19, 6382–6390.

524 Schouten, S., Hopmans, E.C., Schefuß, E., Sinninghe Damsté, J.S., 2002. Distributional
 525 variations in marine crenarchaeotal membrane lipids: a new organic proxy for
 526 reconstructing ancient sea water temperatures? *Earth and Planetary Science Letters*
 527 204, 265–274.

528 Schouten, S., Hudeg, C., Hopmans, E.C., Kienhuis, M.V.M., Sinninghe Damsté, J.S.,
 529 2007. Analytical methodology for TEX₈₆ paleothermometry by high performance
 530 liquid chromatography/atmospheric pressure chemical ionization-mass spectrometry.
 531 *Analytical Chemistry* 79, 2940–2944.

532 Seki, O., Schmidt, D.N., Schouten, S., Hopmans, E.C., Sinninghe Damsté, J.S.,
 533 Pancost, R.D., 2012. Paleooceanographic changes in the eastern equatorial Pacific over
 534 the last 10 Myr. *Paleoceanography* 27, PA3224.

535 Spero, H.J., Lea, D.W., 2002. The cause of carbon isotope minimum events on glacial
 536 terminations. *Science* 296, 522–525.

537 Steger, J.M., Collins, C.A., Chu P.C., 1998. Circulation in the Archipelago de Colon
 538 (Galápagos Islands). *Deep-Sea Research II* 45, 1093–1114.

539 Strub, P.T., Mesias, J.M., Montecino, V., Rutllant, J., Salinas, S., 1998. Coastal ocean
 540 circulation off western South America. In: Robinson, A.R., Brink, K.H. (Eds.), *The*
 541 *Sea, Ideas and Observations on Progress in the Study of the Seas*, Vol. 11, Global
 542 *Coastal Ocean Regional Studies and Syntheses*. Wiley, New York, pp. 273–313.

543 Talley, L.D., Pickard, G.L., Emery, W.J., Swift, J.H., 2011. Descriptive physical
 544 oceanography. 6th Edition. Elsevier, Boston.
 545 Toggweiler, J.R., Russell, J.L., Carson, S., 2006. Midlatitude westerlies, atmospheric
 546 CO₂, and climate change during the ice ages. *Paleoceanography* 21, PA2005.
 547 Trenberth, K.E., 1997. The definition of El Niño. *Bulletin of the American*
 548 *Meteorological Society* 78, 2771–2777.
 549 Turich, C., Freeman, K.H., Bruns, M.A., Conte, M., Jones, A.D. and Wakeham, S.G.,
 550 2007. Lipids of marine Archaea: Patterns and provenance in the water-column and
 551 sediments. *Geochimica et Cosmochimica Acta* 71, 3272–3291.
 552 Waliser, D.E., Gautier, C., 1993, A satellite-derived climatology of the ITCZ. *Journal of*
 553 *Climate* 6, 2162–2174.
 554 Wang, C., Fiedler, P.C., 2006. ENSO variability in the eastern tropical Pacific: a review.
 555 *Progress in Oceanography* 69 (2–4), 239–266.
 556 Weijers, J.W.H., Schouten, S., Spaargaren, O.C., Sinninghe Damsté, J.S., 2006.
 557 Occurrence and distribution of tetraether membrane in soils: implications for the use
 558 of the BIT index and the TEX₈₆ SST proxy. *Organic Geochemistry* 37, 1680–1693.
 559 Wyrski, K., 1967. Circulation and water masses in the eastern equatorial Pacific Ocean.
 560 *International Journal of Oceanology and Limnology* 1, 117–147.
 561 Wyrski, K., 1975. El Niño – The dynamic response of the equatorial Pacific Ocean to
 562 atmospheric forcing. *Journal of Physical Oceanography* 5, 572–584.
 563 Wyrski, K., 1981. An estimate of equatorial upwelling in the Pacific. *Journal of Physical*
 564 *Oceanography* 11, 1205–1214.

- Xie, R.C., Marcantonio, F., 2012. Deglacial dust provenance changes in the eastern equatorial Pacific and implications for ITCZ movement. *Earth and Planetary Science Letters* 317-318, 386–395.
- Yamamoto, M., Shiraiwa, Y., Inouye, I., 2000. Physiological responses of lipids in *Emiliania huxleyi* and *Gephyrocapsa oceanica* (Haptophyceae) to growth status and their implications for alkenone paleothermometry. *Organic Geochemistry* 31, 799–811.
- Yamamoto, M., Yamamuro, M., Tanaka, Y., 2007. The California current system during the last 136,000 years: response of the North Pacific High to precessional forcing. *Quaternary Science Reviews* 26, 405–414.
- Yamamoto, M., Okino, T., Sugisaki, S., Sakamoto, T., 2008. Late Pleistocene changes in terrestrial biomarkers in sediments from the central Arctic Ocean. *Organic Geochemistry* 39, 754–763.
- Yamamoto, M., Shimamoto, A., Fukuhara, T., Tanaka, Y., Ishizaka, J., 2012. Glycerol dialkyl glycerol tetraethers and TEX₈₆ index in sinking particles in the western North Pacific. *Organic Geochemistry* 53, 52–62.
- Zhang, Y., Sintes, E., Chen, J., Zhnag, Y., Dai, M., Jiao, N., Herndl, G.J., 2009. Role of mesoscale cyclonic eddies in the distribution and activity of Archaea and bacteria in the South China Sea. *Aquatic Microbial Ecology* 56, 65–79.

585 Figure

586

587 Figure 1. Map showing mean annual SST (Locarnini et al., 2010), the location of ODP
 588 Site 1239 (this study), Sites 1240, ME24, ME27, N22P, CD38-02 and the surface and
 589 subsurface ocean currents in the EEP. SEC = South Equatorial Current, NEC = North
 590 Equatorial Current, EUC = Equatorial Undercurrent, NECC = North Equatorial
 591 Countercurrent; modified after Kessler (2006) and Pennington et al. (2006).

592

593

594

595 Figure 2. Records of (a) Niño 3.4 index (averaged SST anomaly in the region 170°W–
 596 120°W, 5°S–5°N) (Trenberth, 1997) and monthly SST at 0.5°S, 82.5°W (Reynolds et al.,
 597 2002) from January 1982 to December 2011 (see online data at
 598 [http://iridl.ldeo.columbia.edu/SOURCES/.NOAA/.NCEP/.EMC/.CMB/.GLOBAL/.Reyn](http://iridl.ldeo.columbia.edu/SOURCES/.NOAA/.NCEP/.EMC/.CMB/.GLOBAL/.Reyn_SmithOIv2/.monthly/)
 599 [_SmithOIv2/.monthly/](http://iridl.ldeo.columbia.edu/SOURCES/.NOAA/.NCEP/.EMC/.CMB/.GLOBAL/.Reyn_SmithOIv2/.monthly/) for detail); (b) Seasonal vertical water structure at 0.5°S, 82.5°W
 600 (see online data at
 601 <http://iridl.ldeo.columbia.edu/SOURCES/.LEVITUS94/.MONTHLY/.temp/>)

Figure 3. Variations in (a) $\text{TEX}_{86}^{\text{H}}$ - and $\text{U}^{\text{K}}_{37'}$ -derived temperatures; (b) the concentration of alkenones and isoprenoid GDGTs; (c) $\delta^{18}\text{O}$ of the benthic foraminifera *C. wuellerstorfi* at the study site (Rincon-Martinez et al., 2010); (d) branched isoprenoid tetraether (BIT) index.

Figure 4. Conceptual model of ΔT in upwelling and non-upwelling conditions.

609

610

611

612 Figure 5. Variation in (a) the difference between $\text{TEX}_{86}^{\text{H}}$ and U_{37}^{K} temperatures (ΔT)
 613 and the abundance ratio of GDGTs to alkenones (GDGT/alkenone); (b) organic carbon
 614 content at ME0005A-24JC and ME0005A-27JC (Dubois and Kienast, 2011); (c) $\delta^{15}\text{N}$ of
 615 bulk sediments from cores CD38-02 and NH22P at the Peru and Mexican margins,
 616 respectively (Ganeshram et al., 2000) and core ME0005A-27JC in the EEP (Dubois and
 617 Kienast, 2011); (d) $\text{DT}-\delta^{18}\text{O}_{\text{sw}}$ at ODP Site 1240 (Pena et al., 2008); (e) Mg/Ca-derived
 618 temperature difference (smoothed; west-east) between the western equatorial Pacific

619 (ODP Site 806B; Medina-Elizalde and Lea, 2005) and the EEP (core TR163-19; Lea et
 620 al., 2000).

621

622

623 Figure 6. Ocean and atmospheric conditions in the eastern Pacific region during
 624 deglaciations. The weaker North Pacific High and the stronger South Pacific High
 625 resulted in the northward shift of the ITCZ, the southward shift of the southern westerlies
 626 and the intensification of upwelling in the EEP, the Peru margin and the Southern Ocean.

Appendix I. Structure of glycerol dialkyl glycerol tetraethers