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2	terminations
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13	

14 Abstract

 TEX_{86}^{H} - and U_{37}^{K} -derived paleotemperatures, and isoprenoid glycerol dialkyl glycerol 15 16 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 17 between TEX₈₆^H- and U^K₃₇-derived temperatures (Δ T) and the abundance ratio of 18 19 GDGTs to alkenones (GDGT/alkenone ratio) are potential upwelling indices which show 20 a consistent results with other upwelling indices. The ΔT and GDGT/alkenone ratio were maximal during the last five deglaciations, suggesting intensified upwelling. The 21 22 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 23

Hemisphere atmospheric circulation induced the intensification of the subtropical highpressure 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.

28

29 **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 δ^{18} 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 TEX_{86}^{H} and U_{37}^{K} for Ocean Drilling Program (ODP) Site 1239 and interpret the U_{37}^{K} and TEX_{86}^{H} records for the last 430,000 years. On the basis of this interpretation, we propose the difference between TEX_{86}^{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.

57

58 Modern physical oceanography

59 The zonal surface current system in the eastern tropical Pacific (ETP) consists of 60 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; 61 3°N to 10°S). The SEC originates as a combination of the waters from the North 62 63 Equatorial Counter Current (NECC), the Equatorial Undercurrent (EUC), and the 64 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 65 of the equator. The NECC, an eastward current flows just north of the equator and is 66 67 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 68 69 NECC there is a narrow equatorial front (EF) that separates warm low-salinity waters in

70 the north from cool high-salinity waters in the south (Fig. 1; Strub et al., 1998). This front 71 is observable from July to September at about 2.5°N with a strong meridional SST 72 gradient. In contrast, the EF position is unclear from January to April, when the southeast 73 trades winds collapse and SST south of the Equator increases owing to reduced upwelling. 74 The condition of the EF is correlated with the displacement of the intertropical 75 convergence zone (ITCZ) (e.g., Pak and Zaneveld, 1974; Chelton et al., 2001; Raymond 76 et al., 2004). The ITCZ reaches its northernmost extent in the month of August (~12°N) 77 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). 78

79 The most influential subsurface current in this region is the EUC that flows 80 eastward beneath the SEC. The EUC is fed by the saline New Guinea Coastal Undercurrent at its the western boundary (Talley et al., 2011) and flows within the 81 82 equatorial thermocline and shoal as it approaches the Galapagos Islands (Kessler, 2006). 83 When it reaches the Galapagos Islands, it splits into two branches (Steger et al., 1998) 84 with the main branch flowing southward to merge with the Peruvian Undercurrent, which 85 provides a for source of the Peru coastal upwelling (Brink et al., 1983), the other branch 86 continues to flow eastward, merging with the NECC (Wyrtki, 1967; Fieldler and Tally, 87 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 (Wrytki, 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).

106

107 **2. Materials and methods**

108 ODP Site 1239 (0°40.32' S, 82°4.86' W; 1414 m water depth) is located near the 109 eastern crest of the Carnegie Ridge and ~120 km off the coast of Ecuador (Fig. 1). The 110 sediments are dominated by light to dark olive gray foraminifera-nannofossil ooze with 111 varying amounts of diatoms, clay, and micrite (Mix et al., 2003). The age-depth model of 112 this core was established by Rincon-Martinez et al. (2010) based on correlation of the δ^{18} O record of the benthic foraminifera *Cibicidoides wuellerstorfi* with the LR04 global 113 stack (Lisiecki and Raymo, 2005). In total, 236 samples were taken from 0.02 to 14.73 114 115 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).

123 The F3 fraction was analyzed with a Hewlett-Packard Model 6890 gas 124 chromatograph with on-column injection and electronic pressure control inlet systems 125 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 (60 126 m; i.d., 0.25 mm; thickness, 0.25 µm). The oven temperature was programmed from 70 to 127 290°C at 20°C min⁻¹, from 290 to 310°C at 0.5°C min⁻¹, and then held for 30 min. 128 129 Quantification of di- and tri-unsaturated C₃₇ alkenones were achieved by comparing the peak areas with that of an internal standard $(n-C_{36}H_{74})$ on the gas chromatogram. 130

131 The alkenone unsaturation index U_{37}^{K} was computed from the concentrations of 132 di-unsaturated (C_{37:2}MK) and tri-unsaturated (C_{37:3}MK) alkenones using the following 133 equation by Prahl et al. (1988):

134

135
$$U_{37'}^{K} = [C_{37:2}MK] / ([C_{37:3}MK] + [C_{37:2}MK])$$

136

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

- 140 $U_{37}^{K} = 0.034T + 0.039$
- 141

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

143

144 An aliquot of F4 was dissolved in hexane-2-propanol (99/1, v/v). GDGTs were 145 analyzed using high-performance liquid chromatography-mass spectrometry (HPLC-MS) 146 with an Agilent 1100 HPLC system connected to a Bruker Daltonics micrOTOF-HS time-147 of-flight mass spectrometer. Separation was conducted using a Prevail Cyano column (2.1 148 x 150 mm, 3µm; Alltech) maintained at 30°C following the method of Hopmans et al. 149 (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 150 151 500–1500). Compounds were identified by comparing mass spectra and retention times 152 with those of GDGT standards (formed from the main phospholipids of Thermoplasma 153 acidophilum via acid hydrolysis).

154 Quantification was achieved by integrating the summed-peak areas in the 155 $(M+H)^+$ and the isotopic $(M+H+1)^+$ ion traces and comparing these to the peak area of an internal standard (C_{46} GDGT) in the (M+H)⁺ ion trace, following to the method of Huguet 156 157 et al. (2006). The correction value of ionization efficiency between GDGTs and the 158 internal standard was obtained by comparing the peak areas of T. acidophilum-derived 159 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 TEX_{86}^{H} 160 161 (applicable in warm water) were calculated from the concentrations of GDGT-1, GDGT-2,

162	GDGT-3, and a regioisomer of crenarchaeol using the following expressions (Schouten et
163	al., 2002; Kim et al., 2010):
164	
165	TEX ₈₆ = ([GDGT-2] + [GDGT-3] + [Crenarchaeol regioisomer]) /
166	([GDGT-1] + [GDGT-2] + [GDGT-3] + [Crenarchaeol regioisomer])
167	$\text{TEX}_{86}^{H} = \log (\text{TEX}_{86})$
168	
169	Temperature was calculated according to the following equation based on a global core-
170	top calibration (Kim et al., 2010):
171	
172	$T = 68.4TEX_{86}^{H} + 38.6$ (when $T > 15^{\circ}C$)
173	
174	where T = temperature (°C). The analytical accuracy was 0.45 °C in our laboratory.
175	
176	3. Results
177	3.1 GDGTs and TEX_{86}
178	The isoprenoid GDGTs detected in ODP 1239 sediments consist of caldrachaeol
179	(GDGT-0), GDGT-1, GDGT-2, GDGT-3, crenarchaeol, and its regioisomer (Appendix I).
180	The total concentration of isoprenoid GDGTs in sediment varied between 0.6 and 12.8
181	$\mu g.g^{\text{-1}}$ with an average of 5.81 $\mu g.g^{\text{-1}}$ (Fig. 3b). The relative abundances of different
182	isoprenoid GDGTs were nearly uniform with range of 37–54% for crenarchaeol, 26–35%
183	for caldarchaeol and 15–35% for others.

The $\text{TEX}_{86}^{\text{H}}$ -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). $\text{TEX}_{86}^{\text{H}}$ -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 TEX_{86}^{H} derived temperatures, but this concern was not relevant for the samples used in this study.

193

194 3.2. Alkenones and U_{37}^{K}

The total concentration of C_{37} - C_{39} 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).

203

4. Discussion

205 **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^K₃₇'-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^K₃₇'-derived temperature was a maximum of 5.5°C higher than the TEX₈₆^H-derived temperature (Fig. 3a).

Dubois et al. (2009) and Kienast et al. (2012) assumed that U_{37}^{K} reflects mean 211 annual SST because the U_{37}^{K} '-derived temperature in EEP core-top sediments 212 213 corresponded to mean annual SST. A sediment trap study at two sites in the central 214 tropical Pacific showed no significant difference in the sinking flux of alkenone producers (Emiliania huxleyi and Gephyrocapsa oceanica) between strong and weak El 215 216 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 217 218 study site.

219 The behavior of Thaumarchaeota and the production of GDGTs are not fully clear in the EEP. Thaumarchaeota (GDGTs producer) are ubiquitous and abundant throughout 220 221 the seawater column (e.g., Massana et al., 2000; Karner et al., 2001). In the central 222 equatorial Pacific, GDGTs are mainly produced in the thermocline layer (TL) (Turich et al., 2007). Recent case studies assumed that the TEX₈₆^H-derived temperatures in EEP 223 224 sediments reflect the temperature of the thermocline (30-50 m) rather than SSTs (Ho et 225 al., 2011; Seki et al., 2012). Thaumarchaeota in marine environments have been recognized to be both heterotrophs (e.g., Ouverney and Fuhrman, 2000; Agogué et al., 226 2008; Zhang et al., 2009) and chemoautotrophic nitrifiers (e.g., Könneke et al., 2005; 227 228 Hallam et al., 2006). Organic matter and NH₃ are produced by phytoplankton and by the 229 decay of organic matter in surface and subsurface water, which explains why 230 Thaumarchaeota are produced in both the surface mixed layer (SML) and TL. We thus assume that TEX_{86}^{H} reflects a mixed temperature signal from the SML and TL (Fig.4). 231 232 The production of Thaumarchaeota is fueled by the supply of organic matter and NH₃. 233 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 234 bloom in the mid-latitude northwestern Pacific. GDGT abundance thus may reflect 235 236 primary production and upwelling intensity.

 TEX_{86}^{H} showed higher temperatures than U_{37}^{K} during the some deglaciations 237 (Fig. 3a), but this does not mean that the integrated SST of the SML and TL was higher 238 than the SST of the SML. The calibration of TEX_{86}^{H} to SST was conducted by comparing 239 core-top TEX₈₆^H with mean annual SST (Kim et al., 2010). If the phenomenon of TEX₈₆^H 240 241 recording both the SST and thermocline temperatures is common in tropical oceans, then calibration requires a comparison between core top TEX₈₆^H and integrated temperatures 242 243 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 TEX_{86}^{H} -244 245 derived temperature.

246

247 **4.2. GDGT/alkenone ratio and \Delta T 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 NH₃ and organic matter. In contrast, when upwelling weakens, GDGT production decreases. The GDGT/alkenone ratio can thus be used as an index ofupwelling intensity.

The difference between $\text{TEX}_{86}^{\text{H}}$ - and U_{37}^{K} -derived temperatures (ΔT) was 254 computed by subtracting U_{37}^{K} -derived SST from TEX₈₆^H-derived temperature (ΔT = 255 $\text{TEX}_{86}^{\text{H}} - \text{U}_{37}^{\text{K}}$). U_{37}^{K} reflects the temperature of the SML and $\text{TEX}_{86}^{\text{H}}$ reflects 256 integrated temperatures from the SML and the TL. When upwelling intensifies, the 257 258 temperature gradient between the SML and TL decreases (Fig. 4), and ΔT shifts in a 259 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 260 261 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, 263 260, 274, 310, 330, 340 and 427 ka. The maxima at 15, 127, 243, 340, and 427 ka 264 correspond to glacial terminations (Fig. 5a). Minimal peaks of ΔT occurred at 33, 86, 179, 265 237, 289, and 386 ka. The variation in ΔT is very similar to that in the GDGT/alkenone 266 ratio, although there are some mismatches in MIS 8 and MIS 11. This correspondence 267 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 $\delta^{18}O$ 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 $\delta^{18}O$ (DT- $\delta^{18}O_{sw}$) 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 275 content during deglaciations were reported from sites P6 (Pedersen, 1983), V19-28 (Lyle 276 et al., 1988), and ME0005A-24JC and 27JC (Kienast et al., 2006) in the EEP (Fig. 5b), 277 suggesting that export production was maximized during the last two deglaciations. The 278 elevated ΔT and GDGT/alkenone ratio at the study site indicate not only the 279 intensification of local upwelling, but also the intensification of regional upwelling 280 associated with thermocline shoaling and enhanced export production in the EEP during 281 deglaciations.

282

283 **4.3. Hydrological evolution in the EEP**

The ΔT record mirrors sedimentary $\delta^{15}N$ records from the Peru margin (Fig. 5c), 284 which have been suggested to reflect the intensity of denitrification regulated by Peruvian 285 coastal upwelling (Ganeshram et al., 2000). The trend in $\delta^{15}N$ at the Peru margin was 286 slightly different from those in the EEP (Dubois and Kienast, 2011) and at the Mexican 287 margin (Ganeshram et al., 2000) (Fig. 5c). The maxima of $\delta^{15}N$ at terminations are 288 significant at the Peru margin but not in the EEP or at the Mexican margin, suggesting 289 that δ^{15} N in the eastern Pacific margin was determined by the denitrification in the Peru 290 margin and modified by local factors (Robinson et al., 2009). The correspondence 291 between ΔT and the Peru margin $\delta^{15}N$ records suggests that the upwelling at the study 292 293 site was closely linked with Peruvian coastal upwelling. The study site is located in a 294 region influenced by the coastal upwelling system (Wrytki, 1981; Pennington et al., 295 2006; Talley et al., 2011). Because the southeast trade winds are a principal agent driving 296 coastal upwelling along the west coast of South American continent (Wrytki, 1981;

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

299 The paleo-position of the ITCZ was approximated using dust fluxes across the 300 equator over the last 30 ka (McGee et al., 2007). The results of that analysis suggest that 301 the ITCZ did not shift southward during the last deglaciation. Xie and Marcantonio 302 (2012) precisely estimated the paleo-position of the ITCZ using neodymium isotopes (\mathcal{E}_{Nd}) derived from transect dust obtained by McGee et al. (2007). The average \mathcal{E}_{Nd} values 303 304 from the last glacial and Holocene show similar gradients throughout the equatorial 305 transect, but the latitudinal gradient was stronger, and a steeper interval was evident 306 during the last deglaciation between 5°N and 7°N. This suggests more northerly mean 307 position of the ITCZ.

308 Yamamoto et al. (2007) reconstructed the intensity of the California Current 309 during the last 150,000 years and showed that the subtropical high-pressure cell in the 310 North Pacific weakened during the last two deglaciations. Lyle et al. (2012) suggested 311 that high precipitation in the Great Basin of the western United States during the last 312 deglaciation was not caused by the southward shift of westerly storms, but instead by the 313 northward transport of moist air masses from the tropical Pacific because of the weaker 314 North Pacific High. This presumes that the northeast trade winds were not intensified 315 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 antiphase variation in the subtropical high-pressure cells of both hemispheres was 320 presumably caused by changes in the heat balance between the hemispheres (Fig. 6).

321 The ENSO model has been applied to understand hydrological evolution of the 322 EEP (e.g., Lea et al., 2000; Koutavas et al., 2002; Koutavas and Lynch-Stieglitz, 2003; 323 Martinez et al., 2003; Pena et al., 2008; Rincon-Martinez et al., 2010). Pena et al. (2008) proposed the deep thermocline seawater $\delta^{18}O$ (DT- $\delta^{18}O_{sw}$) based on *Neogloboquadrina* 324 *dutertrei* δ^{18} O at Site 1240 and suggested that EEP hydrology was characterized by a La 325 Niña-like condition during deglaciations. However, the zonal gradient of SST was 326 327 inconsistent with a La Niña-like state during the last deglaciation (Fig. 5e). The DT- $\delta^{18}O_{sw}$ at ODP Site 1240 showed maximum peaks during deglaciations (Pena et al., 2008), 328 329 but the Mg/Ca-SST, between the western and eastern Pacific did not show a large 330 temperature gradient typical of La Niña (Lea et al., 2000). Also, the weaker North Pacific 331 High evidenced during the last deglaciation (Yamamoto et al., 2007; Lyle et al., 2012) is 332 not consistent with a La Niña-like state; a weaker North Pacific High is typical of the 333 modern El Niño condition (Bogad and Lynn, 2001). We thus suggest that intensified upwelling shown by enhanced $DT-\delta^{18}O_{sw}$ at Site 1240 was not linked to a La Niña-like 334 335 state, and an ENSO analogy cannot to be applied to explain hydrological conditions in 336 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.

347 The intensification of the South Pacific High caused southern westerlies to move 348 poleward and the ITCZ to shift northward during deglaciations (Fig. 6). In response, the 349 center of upwelling moved northward and cold tongue upwelling in the EEP area 350 intensified. The stronger South Pacific High during the last deglaciation caused a drier 351 climate in the Patagonia region of South America (de Porras et al., 2012), and the weaker 352 North Pacific High caused a wetter climate in the Great Basin of the western United 353 States (Lyle et al., 2012). This perspective is useful for understanding the hydrological and climatological evolution of the eastern Pacific region. 354

355

356 **5.** Conclusions

The abundance ratio of GDGTs to alkenone (GDGT/alkenone ratio) and 357 difference between TEX₈₆^H- and U^K₃₇'-derived temperature (ΔT) can be used as 358 359 upwelling indices in the EEP. Our new data show that intensification of upwelling 360 occurred in the EEP at each of the last five glacial terminations. The result suggests that 361 the intensification of upwelling was a common phenomenon in the EEP at glacial 362 terminations. The similar timing of intensified upwelling in the EEP, the Peru margin, 363 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 364 eastern Pacific region during deglaciations. 365

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585 Figure





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

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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 December 2011 online data to (see at 598 http://iridl.ldeo.columbia.edu/SOURCES/.NOAA/.NCEP/.EMC/.CMB/.GLOBAL/.Reyn 599 _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 U_{37}^{K} '-derived temperatures; (b) the concentration of alkenones and isoprenoid GDGTs; (c) δ^{18} O of the benthic foraminifera *C. wuellerstorfi* at the study site (Rincon-Martinez et al., 2010); (d) branched isoprenoid tetraether (BIT) index.



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

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Figure 5. Variation in (a) the difference between TEX₈₆^H and U^K₃₇' temperatures (Δ T) and the abundance ratio of GDGTs to alkenones (GDGT/alkenone); (b) organic carbon content at ME0005A-24JC and ME0005A-27JC (Dubois and Kienast, 2011); (c) δ¹⁵N of bulk sediments from cores CD38-02 and NH22P at the Peru and Mexican margins, respectively (Ganeshram et al., 2000) and core ME0005A-27JC in the EEP (Dubois and Kienast, 2011); (d) DT-δ¹⁸O_{sw} at ODP Site 1240 (Pena et al., 2008); (e) Mg/Ca-derived 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 et620 al., 2000).

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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.



628 Appendix I. Structure of glycerol dialkyl glycerol tetraethers