
Millennial-scale Atlantic/East Pacific sea surface temperature linkages during the last 100,000 years

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Abstract :

Amplifying both internally generated variability and remote climate signals from the Atlantic Ocean via coupled air-sea instabilities, the eastern tropical Pacific (ETP) is well situated to detect past climate changes and variations in Central American wind systems that dynamically link the Atlantic and the Pacific.

Here we compare new and previously published alkenone-based sea surface temperature (SST) reconstructions from diverse environments within the ETP, i.e. the Eastern Pacific Warm Pool (EPWP), the equatorial and the northern Peruvian Upwelling regions over the past 100,000 yr. Over this time period, a fairly constant meridional temperature gradient across the region is observed, indicating similar hydro-graphic conditions during glacial and interglacial periods. The data further reveal that millennial-scale cold events associated with massive iceberg surges in the North Atlantic (Heinrich events) generate cooling in the ETP from similar to 8 degrees N to similar to 2 degrees S. Data from Heinrich event 1, however, indicate that the response changes sign south of 2 degrees S. These millennial-scale alterations of the SST pattern across diverse environments of the ETP support previous climate modeling experiments that suggested an Atlantic-Pacific connection caused by the intensification of the Central American gap winds, enhanced upwelling and mixing north of the equator and supported by positive air-sea feedbacks in the eastern tropical Pacific.

Highlights

► Millennial-scale cooling events occur in the East Pacific during Heinrich events. ► N. Atlantic SST anomalies affect Central American gap winds and thus the E. Pacific. ► Hydrographic conditions were similar during glacial and interglacial periods.

Keywords : Eastern Pacific, Heinrich events winds sea surface temperature alkenone, EOF

36 **1. Introduction**

37 The ETP is home to the El Niño-Southern Oscillation (ENSO) phenomenon, which
38 affects weather and climate patterns worldwide. Recent studies on remote triggers of ETP
39 climate variability have suggested possible extratropical atmospheric influences
40 (Alexander et al., 2010; Caballero and Anderson, 2009; Vimont et al., 2003), along with
41 potential effects from tropical Atlantic SSTs (Okumura et al., 2009; Timmermann et al.,
42 2007; Zhang and Delworth, 2005). Given the limited degrees of freedom in the short
43 instrumental record, the Atlantic/tropical Pacific linkage is difficult to establish with
44 statistical confidence. Paleo-climate data could provide additional insight into the
45 mechanisms that communicate Atlantic climate anomalies into the tropical Pacific.
46 Moreover, understanding this pan-oceanic atmospheric bridge and the associated changes
47 in moisture transport across Central America (Leduc et al., 2007; Richter and Xie, 2010)
48 will shed further light on the long-term behavior and stability of the Atlantic Meridional
49 Overturning Circulation vis-à-vis Deep Water Formation in the North Pacific (Okazaki et
50 al., 2010).

51 Here we compare three ETP alkenone-based SST reconstructions of the past 100,000
52 years that document a recurring link between Atlantic and Pacific climate on millennial
53 timescales, thus extending previous studies that identified a connection between Atlantic
54 climate change, Pacific hydroclimate variability (Benway et al., 2006; Leduc et al., 2007;
55 Pahnke et al., 2007) and Pacific SSTs (Kienast et al., 2006; Pahnke et al., 2007; Koutavas
56 and Sachs, 2008) during the last glacial termination. We perform empirical orthogonal
57 function (EOF) analyses of these three 100,000 years long SST records and of a

58 compilation of eleven previously published alkenone-based SST records from the ETP
59 covering the last 25,000 years.

60

61 **2. Regional setting**

62 The ETP is characterized by prominent climatic asymmetries and large seasonal
63 variations in wind patterns, surface currents, temperature and salinity. North of the
64 equator, the EPWP is characterized by annual mean temperatures exceeding 27.5°C and
65 exceptionally low salinities (~32 practical salinity units, psu), resulting from intense
66 rainfall associated with the Intertropical Convergence Zone (ITCZ) and atmospheric
67 water vapor export from the Caribbean across the Panama Isthmus (Joussaume et al.,
68 1986; Mitchell and Wallace, 1992; Li and Philander, 1996; Xie et al., 2005) via the
69 northeasterly trade winds (Fig. 1).

70 Straddling the equator on the other side are the southeasterly trade winds that
71 converge into the northern hemisphere ITCZ. These winds cause Ekman divergence
72 along the equator and give rise to the equatorial Cold Tongue (mean SST of 24°C, Fig.
73 1d). The southeasterly trades also cause upwelling off the coast of Peru. The Peru Current
74 advects cold coastal waters into the far eastern tropical Pacific, thus contributing to the
75 low temperatures in the Cold Tongue region.

76 The seasonal cycle of ETP SST and ocean currents is linked to the seasonal
77 meridional migration of the ITCZ (Fig. 1 a-d). During present-day boreal summer &
78 autumn (May to late November), when the ITCZ is in its northernmost position (10-12°N,
79 Fig. 1b), the southerly winds are strongest over the equator and the Equatorial and
80 Peruvian Upwelling intensify (Fig. 1d; Wyrtki, 1981). A strong meridional SST gradient

81 develops across the equator, between the EPWP and the Cold Tongue with a well-
82 pronounced Equatorial Front near 2–5°N (Wyrski, 1996; Pak and Zaneveld, 1974). At the
83 equator, SSTs are at their minimum in September (Fig. 1d).

84 This stage is contrasted by conditions in boreal winter and spring (December to
85 April), when the ITCZ is at its most equatorial position, strong northeasterly trades cross
86 the Central American Isthmus (Fig. 1a) and southeast trade winds and the equatorial Cold
87 Tongue are relatively weak (Fig. 1c; Li and Philander, 1996). SSTs reach their maximum
88 in March. As the strong northeasterly trades accelerate through topographic gaps in the
89 Central American Cordillera, they form smaller-scale features known as the Tehuantepec,
90 Papagayo and Panama wind jets (Fig. 1a; Xie et al., 2005; Chelton et al., 2000). The wind
91 stress curl associated with these jets causes localized upwelling, thermocline changes and
92 SSTs minima extending off the coast off Central America (Fig. 1a and 1c; Kessler, 2002;
93 Xie et al., 2005; Willett et al., 2006). Of particular importance is the Panama Jet (January
94 to April) which generates a cold SST patch in its wake (Fig. 1c) that can inhibit
95 convection and breaks the winter ITCZ into two parts (Alory et al., 2012). The Papagayo
96 Jet and its related wind stress curl pattern create the so-called Costa Rica Dome, an
97 oceanic upwelling center where the thermocline ascends to very near the sea surface
98 (Fiedler, 2002; Kessler, 2006). A local SST minimum at 9°N, 89°W marks the Costa Rica
99 Dome (Fig.1c), which ranges in diameter from 100 to 900 km. These gap winds (i.e.,
100 winds that are accelerated by an along-gap pressure gradient) have been associated with
101 variations in the Caribbean trade winds (Frankenfield, 1917) and high pressure systems of
102 midlatitude origin that move southeastward across the Gulf of Mexico (a.k.a. “Northers”
103 or “Central American Cold Surges”; Schultz et al., 1997; 1998).

104

105 **3. Material**

106 Here we present an extended 100 kyr-long record of alkenone-based (UK'37-) SST
107 reconstructions from core ME0005A-43JC (hereafter ME43) collected off the Costa
108 Rican margin in the EPWP (Fig. 1a). We compare this record with two UK'37-SST
109 reconstructions from the equatorial Cold Tongue environment (ME0005A-24JC and
110 ME0005A-27JC, hereafter ME24 and ME27, respectively), recently published by Dubois
111 et al. (2011). Note that Kienast et al. (2006) recently published SST reconstructions from
112 the last 25 kyr for ME24 and Dubois et al. (2009) for ME27 and ME43.

113

114 **3.1 Core locations**

115 Core ME43 (7°51.35'N, 83°36.50'W, 1368 m water depth) was collected during the
116 ME0005A expedition aboard the *R/V Melville* in 2000, off the Costa Rican margin in the
117 EPWP. Sediment cores ME24 (0°01'N, 86°27'W, 2941 m water depth) and ME27
118 (1°51'S, 82°47'W, 2203 m water depth) were recovered during the same expedition as
119 core ME43. The strategic location of these cores across this complex oceanographic
120 region is ideal to detect hydrographic changes (Fig. 1). As shown in the main text, both
121 coastal and equatorial upwelling systems co-exist in the ETP. Our southernmost core site
122 ME27 is affected partly by the highly productive waters originating from of the coastal
123 upwelling region off Peru, where deep subantarctic-sourced, nutrient-rich water are
124 upwelling that originate from the deeper part of the Equatorial Under Current (EUC).
125 Core ME24 is situated within the Equatorial Upwelling zone, where shallower waters
126 from the upper EUC branch reach the surface via the equatorial divergence (Dugdale et

127 al., 2004). The local hydrology at our northern most core site ME43 is not affected by
128 local upwelling dynamics, as it is situated in the lee of the Talamanca Cordillera, which
129 reaches altitudes of 3,000 to 4,000 m above sea level (Chelton et al., 2000). At this core
130 location, mean annual SST is higher than 28°C (Benway et al., 2006). However, site
131 ME43 is situated in-between the Costa Rica Dome and the SST minimum associated with
132 the Panama Jet (Fig. 1c), and may be sensitive to intensification of either of these
133 upwelling centers.

134

135 **3.2 Age models**

136 All age models are based on previously published planktonic foraminifera radiocarbon
137 dates collected via accelerator mass spectrometry (AMS), benthic oxygen isotope
138 stratigraphy, and the detection of the Los Chocoyos Ash Layer (Drexler et al., 1980) in
139 ME24 and ME27 (see Tables 1-2 and Fig. 2 for details and data sources). We updated the
140 age models for all three cores using the recently published MARINE13 calibration data
141 set (Reimer et al., 2013). All radiocarbon dates were converted to calendar ages using the
142 Calib 7.0 program. For ME43, a reservoir age of 558 years, the regional average for the
143 west coast of Central America (Stuiver and Reimer, 1993) was employed. For ME24 and
144 ME27, a reservoir age of 467 years was employed. The calendar age with the median
145 probability was selected (Table 1). Age models were created by linearly interpolating
146 between derived calendar ages.

147 Age model for core ME43 was updated from Benway et al. (2006). The use of a new
148 calibration data set (MARINE13) only leads to small deviations (<450 yr) from the
149 calendar ages published by Benway et al. (2006) using MARINE04 (Hughen et al., 2004;

150 See Table 1). Note that for the interval 23-85 kyr BP, Benway et al. (2006) correlated 3
151 planktonic $\delta^{18}\text{O}$ tie points to GISP2 $\delta^{18}\text{O}$ (Grootes and Stuiver, 1997) and 6 benthic $\delta^{18}\text{O}$
152 tie points to MD95-2042, a core from the Iberian Margin (MD95-2042, 37°48'N,
153 10°10'W) whose stratigraphy was based on correlating planktonic $\delta^{18}\text{O}$ to GRIP $\delta^{18}\text{O}$
154 (Shackleton et al., 2000). Here we chose to correlate all three benthic oxygen isotope
155 records (A. Mix, unpubl. data) to the benthic stack LR04 (Lisiecki and Raymo, 2005; See
156 Fig. 2). This leads to small changes in the original ME43 age model, reaching a
157 maximum of 1.4 kyr around 46 and 54 kyr BP (Fig. 2 and Table 2). These results are
158 available from the National Climatic Data Center at [http://www.ncdc.noaa.gov/data-](http://www.ncdc.noaa.gov/data-access/paleoclimatology-data/datasets/paleoceanography)
159 [access/paleoclimatology-data/datasets/paleoceanography](http://www.ncdc.noaa.gov/data-access/paleoclimatology-data/datasets/paleoceanography).

160 As pointed out in previous studies, the assumption of a constant reservoir age in
161 radiocarbon dating is challenging, especially in upwelling regions, given the likelihood of
162 variations in upwelling intensity and exchange with the atmosphere (Marchitto et al.,
163 2007; Singarayer et al., 2008). While we acknowledge the potential chronological error
164 associated with this uncertainty, we do not yet have the ability to account for this.

165 The temporal resolution of core ME43 is one sample every 400 years on average for the
166 last 30 kyr, and 1 sample every 610 years on average for the interval 30-100 kyr BP.
167 Cores ME24 and ME27 have similar resolution with - on average - one sample every 340
168 years and one sample every 650 years, respectively.

169

170 **4. Methods**

171 **4.1 Laboratory analysis**

172 Core ME43 was analyzed for alkenone unsaturation at Dalhousie University. Total lipids

173 were extracted from 1.5 to 3 g aliquots of freeze-dried sediment samples using a Dionex
174 Accelerated Solvent Extraction system (ASE200). These extracts were saponified using
175 potassium hydroxide and purified through silica column chromatography. The purified
176 extracts were analyzed by gas chromatography (Agilent 6890N) at Dalhousie. The UK'37
177 index was calculated according to the relative concentrations of the di- and tri-
178 unsaturated C37 alkenones: $UK'37 = [C37:2]/[C37:2 + C37:3]$. Sea-surface temperature
179 estimates (UK'37-SST) were calculated from the ratio of the concentration of the di- and
180 triunsaturated alkenones, using the calibration of Prahl et al. (1988): $SST (^{\circ}C) = (UK'37 -$
181 $0.039)/0.034$. This calibration has recently been shown to best relate alkenone
182 unsaturation to sea surface temperatures in the ETP (Kienast et al., 2012). Based on
183 replicate analysis, the analytical precision of the method is ± 0.01 UK'37 units ($0.3^{\circ}C$).
184 The standard calibration error is estimated at $\pm 1.5^{\circ}C$. These results are available from the
185 National Climatic Data Center at [http://www.ncdc.noaa.gov/data-](http://www.ncdc.noaa.gov/data-access/paleoclimatology-data/datasets/paleoceanography)
186 [access/paleoclimatology-data/datasets/paleoceanography](http://www.ncdc.noaa.gov/data-access/paleoclimatology-data/datasets/paleoceanography).

187

188 **4.2 Empirical orthogonal functions**

189 Empirical orthogonal function (EOF) analyses are performed here on two different
190 compilations of UK'37-SST reconstructions of the ETP, abbreviated as vector
191 $\mathbf{T}(t)=(T_i(t))$. The first one is based on the SST reconstructions of cores ME43, ME24 and
192 ME27 and covers the period from 100 to 5 kyr BP. The second one is performed on a set
193 of recently published alkenone-based SST reconstructions from the ETP and covers the
194 last deglacial period (25-5 kyr BP).
195 EOF analysis is a commonly used statistical method that determines the leading

196 orthogonal patterns (denoted as \mathbf{e}_i) of variability of a multivariate dataset $\mathbf{T}(t)$ and their
197 corresponding principal components $p_i(t)$, such that $\mathbf{T}(t)=\sum_i p_i(t) \mathbf{e}_i$. The corresponding
198 principal component $p_i(t)$ can be expressed as the scalar product $p_i(t)=\mathbf{T}(t)\bullet\mathbf{e}_i$.
199 The contribution of each EOF to the SST evolution at a sediment core location is
200 determined by multiplying the corresponding principal components with the EOF pattern
201 loading at this site. EOF patterns are ranked based on their explained variance to the total
202 multivariate dataset. Prior to the EOF analysis all the SST reconstructions were linearly
203 interpolated onto the same equidistant time grid with a grid spacing of 100 years.
204 Locations of the ETP cores used in the deglacial EOF computation are given in Table 3,
205 along with the actual EOF values for each core. Original UK'37-SST reconstructions
206 used in the deglacial EOF are shown in Fig. 3. Note that we have used an updated age
207 model for core V19-30, as presented by Kienast et al. (2013).

208

209 **5. Results**

210 **5.1 UK'37-SST reconstructions**

211 All three SST records reveal a long-term cooling trend from 100 to 17 kyr BP,
212 followed by a more rapid warming during the deglaciation and the Holocene (Fig. 4b-d).
213 The dominant glacial/interglacial SST magnitude ranges from 1.5°C-2.5°C. A striking
214 feature revealed by all three data sets is an unusually warm period (80-100 kyr BP)
215 during MIS5, with temperatures exceeding those of the Holocene by up to 1°C.

216 To determine the spatial inhomogeneity of orbital-scale variability we calculated the
217 SST differences between the sites (Δ SST, Fig. 4f-h). In order to avoid large amplitude
218 variability due to age model uncertainties and outlier samples, we computed the Δ SST

219 after a 15 points smoothing of the SST curves re-sampled at 500 years intervals (Fig. 4b-
220 d). We find that the low-frequency temperature gradients between the 3 sites remain
221 relatively stable through the last 100 kyr varying by no more than +/- 1°C. A slight long-
222 term increase in the gradient between the EPWP and Cold Tongue can be observed in Fig.
223 4g and h (ME43-ME24 and ME43-ME27).

224 Millennial-scale oscillations are superimposed on this long-term trend, with highest
225 amplitudes during MIS3. Dubois et al. (2011) noticed that the SST difference between
226 ME24 and ME27, both located in the Cold Tongue region remains relatively constant
227 over the last 100 kyr, except during rapid and pronounced millennial-scale cooling
228 intervals during which the SST gradient is minimal. Interestingly, the SST record of
229 ME43, located further north in the EPWP, reveals millennial-scale cooling intervals that
230 appear to coincide within dating uncertainty with the cooling events observed in the Cold
231 Tongue. Note that the Δ SST records in Fig. 4f-h cannot be used to infer millennial-scale
232 changes between the cores since these records have been smoothed.

233

234 **5.2 EOF analysis**

235 To identify the leading patterns of orbital and millennial-scale variability during the
236 last glacial period, we computed the leading Empirical Orthogonal Function (“long-term”
237 EOF) mode of the linearly interpolated SST records ME24, ME27 and ME43. The first
238 mode explains 93.7% of the joint variance for these three cores and shows a monopole
239 pattern of EOF loadings (Fig. 5a). The corresponding principal component (Fig. 4e, 5d) is
240 characterized by the large-scale glacial signal and millennial-scale variability. Millennial-
241 scale features in this mode coincide with the exceptionally cold stadials in northeastern

242 North Atlantic SST (Fig. 5d, blue line) and the associated periods of iceberg surging
243 (Heinrich events) H1, H2, H4, H5 and H6. H3 is not evident and H4 is not well marked
244 in our Pacific SST records. However, the interpretation of ETP sedimentary records is
245 complicated by the fact that they are of lower temporal resolution and affected by local
246 processes. While there is thus no one-on-one match between North Atlantic/Greenland
247 events and the ETP paleoceanographic events presented here, we nevertheless interpret
248 the cooling in the ETP during H1, H2, H4, H5 and H6 to be indicative of a strong linkage
249 between the Atlantic and the ETP.

250 This Atlantic/Pacific linkage is further corroborated by a deglacial EOF analysis of
251 eleven ETP SST reconstructions covering the period of the last glacial termination and
252 the early Holocene. For this analysis we used UK'37-based SST from cores V19-27,
253 V19-28, V19-30, V21-30, RC11-238, MD02-2529, JPC32-14, ME24, ME27, ME43 and
254 TR31 (See Table 3 for locations and original references). Note that this set of cores is
255 located between 8.1°N and 3.3°S and thus extends further south than the longer cores
256 ME43, 24 and 27. Figure 5 shows the 1st and 2nd deglacial EOF modes (\mathbf{e}_1 , \mathbf{e}_2) and their
257 corresponding principal components ($p_1(t)$, $p_2(t)$). The leading deglacial SST mode in the
258 ETP, explaining 80.5% of the total variance, is also monopole pattern (Fig. 5b). Its time-
259 evolution is characterized by a cooling during Heinrich event 1 (H1, 17-15 kyr BP), a
260 relatively fast warming at around 15.5 kyr BP, a stalling of temperatures during the
261 Younger Dryas (YD) period and a subsequent gradual temperature rise into the mid
262 Holocene (Fig. 5e). The second EOF mode of deglacial temperatures in the ETP explains
263 10.0% of the variance and reveals a N-S dipole pattern (Fig. 5c). Its time-evolution is
264 characterized by cooling during both H1 and H2 (Fig. 5f).

265

266 **6. Discussion**

267 **6.1 Orbital-scale variations in ETP SST**

268 As shown in Figure 4, the temperature gradients between the 3 long-term records remain
269 relatively stable throughout the last 100 kyr. This near-stationary temperature offset
270 between the EPWP, the Equatorial Upwelling and the northern Peru Upwelling regions
271 over the last 100 kyrs suggests that the dynamics regulating SST on glacial-interglacial
272 timescales affected these environments uniformly. However, a slight long-term increase
273 in the gradient between the EPWP (ME43) and Cold Tongue (ME27 and ME24) can be
274 observed in Fig. 4f and 5g (between 100 and 15 kyr BP). Furthermore, the larger glacial
275 (MIS2-4) gradient between the two Cold Tongue sites (Fig. 4h) suggests a stronger
276 cooling in the Peru Upwelling (ME27) relative to the Equatorial Upwelling environment.
277 This observation supports previous findings of enhanced advection in the Peru Current
278 during glacials compared to interglacials (Kaiser et al., 2005; Dubois et al., 2009).
279 The leading deglacial principal component (Fig. 5b and e) exhibits a monopole pattern,
280 most probably controlled simultaneously by different local and remote forcings. Figure
281 5e suggests that although CO₂ is already increasing at 18 kyr BP (blue curve in Fig. 5e),
282 PC1 does not increase until 15.5 kyr BP partly because it is compensated for by the drop
283 in solar insolation in the boreal fall (October, November, December, OND, orange
284 dashed curve in Fig. 5e) and the resulting temperature effect in December, January,
285 February (DJF). Note: SSTs in the ETP lag the annual cycle insolation forcing and its
286 modulation through the Milankovitch cycles by about 2 months. This lag emerges from
287 the mean mixed layer depth of about 30m and a thermal damping timescale of about 3 –

288 4 months (Barnett et al., 1991).

289

290 **6.2 Millennial-scale variability in the ETP**

291 Most research on suborbital timescales in the Eastern Pacific focused on the
292 hydrologic cycle and sea surface salinity variations (e.g. Benway et al., 2006; Leduc et al.,
293 2007; Pahnke et al., 2007). Significant disparities emerged and were already discussed by
294 Pahnke et al. (2007) and Prange et al. (2010), highlighting the difficulty of paleosalinity
295 reconstruction as well as the complexity of the regional hydrology.

296 Although many SST reconstructions exist for the ETP, using various proxies such as
297 planktonic $\delta^{18}\text{O}$, microfossil assemblages or Mg/Ca, few have the resolution necessary to
298 investigate millennial-scale features. Those that have the required resolution focused on
299 the deglaciation, in particular H1 and the YD (Kienast et al., 2006; Pahnke et al., 2007;
300 Koutavas and Sachs, 2008; Kienast et al., 2013), and will be discussed further in
301 paragraph 6.3. Leduc et al. (2007) presented a UK'37-SST record for core MD02-2529
302 ($8^{\circ}12.33'\text{N}$, $84^{\circ}07.32'\text{W}$; hereafter MD29), which is located close to ME43. The UK'37-
303 SST record of MD29 shows very similar millennial-scale features to the record of ME43,
304 including the cooling events which we observe further south in the Cold Tongue (Fig. 6).
305 Match between the two SST records could be improved even further by carefully aligning
306 the age models, which is beyond the scope of the present manuscript. Note that in this
307 figure, we have applied the Prahl et al. (1988) calibration equation to convert the UK'37
308 index into SST for both of the records.

309 Interestingly, Mg/Ca records of high enough resolution do not reveal millennial-scale
310 events typical of the Northern or Southern Hemisphere (e.g. Lea et al., 2006; Pena et al.,

311 2008). One explanation for these discrepancies might be the fact that Mg/Ca records are
312 markedly affected by the preservation of foraminiferal shells (Herbert, 2003; Mekik et al.,
313 2007; Regenberg et al., in press). Another interpretation may be related to potential
314 seasonal sensitivities of Mg/Ca data in the ETP (Sachs, 2008; Timmermann and Timm,
315 2008) and the fact that in Coupled General Circulation Models the Atlantic/Pacific
316 atmospheric connections is most pronounced during boreal spring (Xie et al. 2008, their
317 Figure 11).

318 The first long-term (Fig. 5d) and the second deglacial (Fig. 5f) principal components
319 exhibit the key millennial-scale features captured by the northeastern Atlantic UK'37-
320 SST reconstruction of core MD01-2444 (37°33.68'N, 10°08.53'W; Martrat et al., 2007)
321 (Fig 5f, blue line), thus supporting the evidence for an Atlantic/EEP connection. MD01-
322 2444, located off the Iberian Margin, has been chosen here as a template for North
323 Atlantic SST variability representing millennial-scale variability with high enough
324 resolution. Nearby SST records from e.g. the Alboran Sea (Cacho et al., 1999) show very
325 similar features to MD01-2444. We think that anomalously strong northeasterly winds
326 across Central America are the most likely explanation for this Atlantic-Pacific linkage,
327 causing the millennial-scale SST excursions we observe in the ETP. Strong cross-isthmus
328 winds (section 2) would enhance upwelling along the Central American Coast and the
329 southwestward advection of cold water towards the equator.

330

331 **6.3 Comparison to modeling studies**

332 The prevalent millennial-scale Atlantic/Pacific connection identified here in proxy
333 SST data is also simulated by Coupled General Circulation Models (CGCMs).

334 Anomalous northeasterly winds across Central America are a feature seen in most
335 CGCMs in response to North Atlantic water hosing experiments (Zhang and Delworth,
336 2005; Timmermann et al., 2007). These experiments typically show a robust
337 intensification of the gap winds across Central America (Xie et al., 2008), an anomalous
338 anticyclone over the Caribbean, a southward shift of the ITCZ in the tropical Atlantic and
339 ETP and an intensification of the Aleutian Low (Okumura et al., 2009). Enhanced
340 northeasterly trades in the EPWP generate enhanced evaporation, mixing and upwelling
341 and subsequent cooling. This cooling can further spread across the equator via air-sea
342 interactions (Xie et al., 2008), such as the Wind-Evaporation-SST feedback, thus
343 explaining the synchronized temperature changes on millennial timescales seen by the 3
344 100 kyr long ETP UK'37-SST data sets presented here.

345 Further south, this surface wind anomaly is responsible for anomalous ocean
346 downwelling. Zhang and Delworth [2005] simulated a cooling of $\sim 1^{\circ}\text{C}$ north of the
347 equator off the coast of Central America and a warming of $\sim 0.4^{\circ}\text{C}$ centered around 7°S in
348 the Cold Tongue. Kienast et al. (2013) recently observed such a pattern of SST
349 distribution during the H1 interval, in particular a slight warming in cores V19-28, V19-
350 30 and TR31, which are located south of our three long records (between $2^{\circ}22'\text{S}$ and
351 $3^{\circ}35'\text{S}$, Fig. 5b and c). This dipole pattern is clearly captured in the second deglacial EOF
352 (Fig. 5c). Note that the EOF pattern (positive in the North) has to be multiplied with the
353 negative value of PC2 during Heinrich events to produce the cooling in the North and
354 warming in the South. Results showing the ETP SST difference between strong and weak
355 AMOC simulated by a GFDL CM2.1 idealized waterhosing experiment (Zhang and
356 Delworth 2005, Timmermann et al. 2007) are shown as contours in Fig. 5c.

357 The limit between cooling and warming in the data set presented by Kienast et al.
358 (2013) for H1 is located $\sim 2^{\circ}\text{S}$ (cooling in RC11-238 at $1^{\circ}31'\text{S}$ and warming in V19-28 at
359 $2^{\circ}22'\text{S}$). Assuming that the boundary between the warming/cooling during Heinrich
360 events has stayed more or less constant through time, our southernmost core ME27 at
361 $1^{\circ}51'\text{S}$ would not have experienced the warming predicted by Zhang and Delworth
362 (2005). Unfortunately, none of the published records located further south extends back
363 in time to 100 kyr BP.

364 Because the deglacial set of cores includes cores south of 2°S (V19-28, V19-30 and
365 TR31) experiencing warming during H1, the amplitude of cooling of the deglacial PC1
366 during H1 is significantly damped in comparison to the long-term PC1 (Fig. 5d and e).
367 On the other hand, all the other records north of 2°S experience cooling, as expected (Fig.
368 3). In summary, the deglacial EOF pattern exhibited by ETP UK'37-SST records is
369 consistent with the meridional dipole pattern in the Eastern Pacific simulated by CGCMs
370 in response to North Atlantic AMOC changes (Zhang and Delworth, 2005; Timmerman
371 et al., 2007, Merkel et al. 2010).

372

373 **7. Conclusions**

374 Over the last 100 kyr, we observe a largely constant temperature offset between the
375 EPWP, the Equatorial Upwelling and the northern Peruvian Upwelling region based on
376 three cores stretching from 8°N and 2°S . This suggests that on orbital time-scales, the
377 dynamics regulating SST affected these three environments uniformly. Superimposed on
378 the glacial-interglacial SST pattern at these sites, we observe abrupt cooling events of up
379 to 2 degrees during most Heinrich events. We suggest that these millennial-scale cooling

380 events are related to cooling in the North Atlantic and the resulting intensification of the
381 Central American gap wind system. Regional upwelling associated with the gap winds
382 likely lead to significant contributions of cold waters originating from the coast of Central
383 America to the equatorial region.

384

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389 who kindly made available his unpublished benthic $\delta^{18}\text{O}$ records of all three cores
390 recovered during the 2000 R/V Melville expedition to the EEP. Thanks are due also to Dr.
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392 tropical Pacific. This contribution benefitted from discussions with A. C. Mix and R. F.
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399

400 **References**

401 Alexander, M.A., Vimont, D.J., Chang, P., Scott, J.D., 2010. The Impact of Extratropical
402 Atmospheric Variability on ENSO: Testing the Seasonal Footprinting Mechanism

403 Using Coupled Model Experiments. *Journal of Climate* 23, 2885–2901.

404 Alory, G., Maes, C., Delcroix, T., Reul, N., Illig, S., 2012. Seasonal dynamics of sea
405 surface salinity off Panama: The far Eastern Pacific Fresh Pool. *J. Geophys. Res.*
406 117, C04028.

407 Barnett, T., Latif, M., Kirk, E., Roeckner, E., 1991. On ENSO physics. *J. Clim.* 4, 487–
408 515.

409 Benway, H.M., Mix, A.C., Haley, B.A., Klinkhammer, G.P., 2006. Eastern Pacific Warm
410 Pool paleosalinity and climate variability: 0–30 kyr. *Paleoceanography* 21, PA3008.

411 Caballero, R., Anderson, B.T., 2009. Impact of midlatitude stationary waves on regional
412 Hadley cells and ENSO. *Geophysical Research Letters* 36, L17704.

413 Cacho, I., Grimalt, J. O., Pelejero, C., Canals, M., Sierro, F. J., Flores, J. A., Shackleton,
414 N., 1999. Dansgaard-Oeschger and Heinrich event imprints in Alboran Sea
415 paleotemperatures. *Paleoceanography*, 14, 698–705.

416 Casey, K.S., Brandon, T.B., Cornillon, P., Evans, R., 2010. "The Past, Present and Future
417 of the AVHRR Pathfinder SST Program", in *Oceanography from Space: Revisited*,
418 eds. V. Barale, J.F.R. Gower, and L. Alberotanza, Springer.

419 Chelton, D.B., Freilich, M.H., Esbensen, S.K., 2000. Satellite observations of the wind
420 jets off the Pacific coast of Central America. Part II: Relationships and dynamical
421 considerations. *Mon. Weather Rev.* 128, 2019–2043.

422 Drexler, J.W., Rose, W.I.J., Sparks, R.S.J., Ledbetter, M.T., 1980. The Los Chocoyos
423 Ash, Guatemala: a major stratigraphic marker in middle America and in three ocean
424 basins. *Quaternary Research* 13, 327–345.

425 Dugdale, R. C. Lyle, M., Wilkerson, F.P., Chai, F., Barbert, R.T., Peng, T.-H., 2004.
426 Influence of equatorial diatom processes on Si deposition and atmospheric CO₂
427 cycles at glacial/interglacial timescales. *Paleoceanography* 19, PA3011.

428 Dubois, N., Kienast, M., Kienast, S., Normandeau, C., Calvert, S.E., Herbert, T.D., Mix,
429 A., 2011. Millennial-scale variations in hydrography and biogeochemistry in the
430 Eastern Equatorial Pacific over the last 100 kyr. *Quaternary Science Review* 30,
431 210–223.

432 Dubois, N., Kienast, M., Normandeau, C., Herbert, T.D., 2009. The east equatorial
433 Pacific cold tongue during the last glacial maximum as seen from alkenone
434 paleothermometry. *Paleoceanography* 24, PA4207.

435 Fiedler, P.C., 2002. The annual cycle and biological effects of the Costa Rica Dome.
436 *Deep Sea Res. Part I* 49, 321–338.

437 Frankenfield, H. C. ‘Northers’ of the Canal Zone. *Mon. Wea. Rev.*, 45, 546–550 (1917).

438 Grootes, P. M. & Stuiver, M. Oxygen 18/16 variability in Greenland snow and ice with
439 10⁻³ to 10⁵-year time resolution. *J. Geophys. Res.* **102**, 26,455–426,470 (1997).

440 Herbert, T. D., 2003. Alkenone paleotemperature determinations. *Treatise on*
441 *Geochemistry* 6, 391-432.

442 Hughen, K. A., et al., MARINE04 marine Radiocarbon age calibration, 26-0 ka BP.
443 *Radiocarbon* 46, 1059-1086 (2004).

444 Joussaume, S., Sadourney, R., Vignal, C., 1986. Origin of precipitating water in a
445 numerical simulation of the July climate. *Ocean–Air Inter.* 1, 43–56.

446 Kaiser, J., F. Lamy and D. Hebbeln (2005), A 70-kyr sea surface temperature record off
447 southern Chile (Ocean Drilling Program Site 1233), *Paleoceanography*, 20, PA4009.

448 Kessler, W.S., 2006. The circulation of the eastern tropical Pacific: A review. *Prog.*
449 *Oceanogr.* 69, 181–217.

450 Kessler, W.S., 2002. Mean three-dimensional circulation in the northeast tropical Pacific.
451 *Journal of Physical Oceanography* 32, 2457–2471.

452 Kienast, M., MacIntyre, G., Dubois, N., Higginson, S., Normandeau, C., Chazen, C.R.,
453 2012. Alkenone unsaturation in surface sediments from the eastern equatorial
454 Pacific: Implications for SST reconstructions. *Paleoceanography* 27, PA1210.

455 Kienast, M., Kienast, S.S., Calvert, S.E., Eglinton, T.I., Mollenhauer, G., François, R.,
456 Mix, A.C., 2006. Eastern Pacific cooling and Atlantic overturning circulation
457 during the last deglaciation. *Nature* 443, 846–849.

458 Kienast, S. S., M. Kienast, A. C. Mix, S. E. Calvert and R. Francois (2007), Thorium-230
459 normalized particle flux and sediment focusing in the Panama Basin region during
460 the last 30,000 years., *Paleoceanography*, 22, PA2213.

461 Kienast, S., Friederich, T., Dubois, N., Hill, P. S., Timmermann, A., Mix, A. C., Kienast,
462 M., 2013. Near collapse of the meridional SST gradient in the eastern equatorial
463 Pacific during Heinrich Stadial 1. *Paleoceanography* 28, 1-12.

464 Koutavas, A., Sachs, J.P., 2008. Northern timing of deglaciation in the eastern equatorial
465 Pacific from alkenone paleothermometry. *Paleoceanography* 23, PA4205.

466 Kusch, S., T. I. Eglinton, A. C. Mix and G. Mollenhauer (2010), Timescales of lateral
467 sediment transport in the Panama Basin as revealed by compound-specific
468 radiocarbon ages of alkenones, *Earth and Planetary Science Letters*, 290, 340-350.

469 Lea, D. W., D. K. Pak, C. L. Belanger, H. J. Spero, M. A. Hall and N. J. Shackleton, 2006.
470 Paleoclimate history of Galapagos surface waters over the last 135,000 yr.

471 Quaternary Science Reviews 25, 1152-1167.

472 Leduc, G., Vidal, L., Tachikawa, K., Rostek, F., Sonzogni, C., Beaufort, L., Bard, E.,
473 2007. Moisture transport across Central America as a positive feedback on abrupt
474 climatic changes. *Nature* 445, 908–911.

475 Li, T., Philander, S.G.H., 1996. On the annual cycle of the eastern equatorial Pacific. *J.*
476 *Clim.* 9, 2986–2998.

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

479 Lüthi, D., Le Floch, M., Bereiter, B., Blunier, T., Barnola, J.-M., Siegenthaler, U.
480 Raynaud, D., Jouzel, J., Fischer, H., Kawamura, K., Stocker, T.F., 2008. High-
481 resolution carbon dioxide concentration record 650,000–800,000 years before
482 present. *Nature* 453, 379–382.

483 Marchitto, T. M., Lehman, S. J., Ortiz, J. D., Flueckiger, J., van Geen, A., 2007. Marine
484 radiocarbon evidence for the mechanism of deglacial atmospheric CO_2 rise. *Science*
485 316, 1456–1459.

486 Martrat, B., Grimalt, J.O., Shackleton, N.J., de Abreu, L., Hutterli, M.A. and Stocker,
487 T.F., 2007. Four climate cycles of recurring deep and surface water destabilizations
488 on the Iberian margin. *Science* 317, 502–507.

489 Merkel, U., M. Prange, and M. Schulz (2010), ENSO variability and teleconnections
490 during glacial times, *Quat. Sci. Rev.*, 29, 86-100, 10.1016/j.quascirev.2009.11.006.

491 Mitchell, T.P., Wallace, J.M., 1992. The annual cycle in equatorial convection and sea
492 surface temperature. *J. Clim.* 5, 1140–1156.

493 Okazaki, Y., Timmermann, A., Menviel, L., Harada, N., Abe-Ouchi, A., Chikamoto,

494 M.O., Mouchet, A., Asahi, H., 2010. Deepwater Formation in the North Pacific
495 During the Last Glacial Termination. *Science* 329, 200–204.

496 Okumura, Y.M., Deser, C., Hu, A., Timmermann, A. and Xie, S.P., 2009. North Pacific
497 Climate Response to Freshwater Forcing in the Subarctic North Atlantic: Oceanic
498 and Atmospheric Pathways. *Journal of Climate* 22, 1424–1445.

499 Pahnke, K., Sachs, J.P., Keigwin, L., Timmermann, A., Xie, S.P., 2007. Eastern tropical
500 Pacific hydrologic changes during the past 27,000 years from D/H ratios in
501 alkenones. *Paleoceanography* 22, PA4214.

502 Pak, H., Zaneveld, J.R.V., 1974. Equatorial Front in the Eastern Pacific Ocean. *Journal of*
503 *Physical Oceanography* 4, 570–578.

504 Pena, L. D., I. Cacho, P. Ferretti and M. A. Hall, 2008. El Nino–Southern Oscillation–
505 like variability during glacial terminations and interlatitudinal teleconnections.
506 *Paleoceanography* 23, PA3101.

507 Prah, F.G., Muehlhausen, L.A., Zahnke, D.L., 1988. Further evaluation of longchain
508 alkenones as indicators of paleoceanographic conditions. *Geochim. Cosmochim.*
509 *Acta* 52, 2303–2310.

510 Prange, M., S. Steph, M. Schulz and L. D. Keigwin, 2010. Inferring moisture transport
511 across Central America: Can modern analogs of climate variability help reconcile
512 paleosalinity records? *Quaternary Science Reviews* 29, 1317–1321.

513 Regenberg, M., Regenberg, A., Garbe-Schönberg, D. and D. W. Lea, in press. Global
514 dissolution effects on planktonic foraminiferal Mg/Ca ratios controlled by the
515 calcite-saturation state of bottom waters, *Paleoceanography*.

516 Reimer, P. J., E. Bard, et al. (2013), *IntCal13 and Marine13 Radiocarbon Age Calibration*

517 Curves 0–50,000 Years cal BP, *Radiocarbon*, 55, 1869-1887.

518 Richter, I., Xie, S.P., 2010. Moisture transport from the Atlantic to the Pacific basin and
519 its response to North Atlantic cooling and global warming. *Climate Dynamics* 35,
520 551–566.

521 Sachs, J. P. (2008). Divergent trends in Holocene SSTs derived from alkenones and
522 Mg/Ca in the Equatorial Pacific. AGU Fall Meeting 2008. San Francisco.

523 Schultz, D.M., Bracken, W.E., Bosart, L.F., Hakim, G.J., Bedrick, M.A., Dickinson, M.J.,
524 Tyle, K.R., 1997. The 1993 Superstorm Cold Surge: Frontal Structure, Gap Flow,
525 and Tropical Impact. *Mon. Weather Rev.* 125, 5–39.

526 Schultz, D.M., Bracken, W.E., Bosart, L.F., 1998. Planetary- and Synoptic-Scale
527 Signatures Associated with Central American Cold Surges. *Mon. Weather Rev.* 126,
528 5–27.

529 Shackleton, N. J., Hall, M. A. & Vincent, E. Phase relationships between millennial-scale
530 events 64,000-24,000 years ago. *Paleoceanography* **15**, 565–569 (2000).

531 Singarayer, J.S., Richards, D.A., Ridgwell, A., Valdes, P. J., Austin, W.E.N., Beck, J.W.,
532 2008. An oceanic origin for the increase of atmospheric radiocarbon during the
533 Younger Dryas. *Geophys. Res. Lett.* 35, L14707.

534 Stuiver, M., Reimer, P.J., 1993 Extended ¹⁴C database and revised CALIB radiocarbon
535 calibration program. *Radiocarbon* 35, 215–230.

536 Timmermann, A., Okumura, Y., An, S.-I., Clement, A., Dong, B., Guilyardi, E., Hu, A.,
537 Jungclaus, J.H., Renold, M., Stocker, T.F., Stouffer, R.J., Sutton, R., Xie, S.-P., Yin,
538 J., 2007. The influence of a weakening of the Atlantic meridional overturning
539 circulation on ENSO. *Journal of Climate* 20, 4899–4919.

540 Timmermann, A. and O. Timm (2008). Mechanisms for deglacial temperature rise in the
541 tropical Pacific. AGU Fall Meeting 2008. San Francisco.

542 Vimont, D.J., Wallace, J.M., Battisti, D.S., 2003. The seasonal footprinting mechanism in
543 the Pacific: Implications for ENSO. *Journal of Climate* 16, 2668–2675.

544 Willett, C.S., Leben, R., Lavin, M.F., 2006. Eddies and mesoscale processes in the
545 eastern tropical Pacific: A review. *Progress in Oceanography* 69, 218–238.

546 Wyrski, K., 1981. An estimate of equatorial upwelling in the Pacific. *Journal of Physical*
547 *Oceanography* 11, 1205–1214.

548 Wyrski, K., 1996. Oceanography of the eastern equatorial Pacific Ocean. *Oceanography*
549 *and Marine Biology. An Annual Review* 4, 33–68.

550 Xie, S.P., Okumura, Y., Miyama, T., Timmermann, A., 2008. Influences of Atlantic
551 climate change on the tropical Pacific via the Central American Isthmus. *Journal of*
552 *Climate* 21, 3914–3928.

553 Xie, S.P., Xu, H.M., Kessler, W.S., Nonaka, M., 2005. Air-sea interaction over the
554 eastern Pacific warm pool: Gap winds, thermocline dome, and atmospheric
555 convection. *Journal of Climate* 18, 5–20.

556 Zhang, R., Delworth, T., 2005. Simulated tropical response to a substantial weakening of
557 the Atlantic thermohaline circulation. *Journal of Climate* 18, 1853–1860.

558

559 **Figure captions**

560 **Figure 1.** (a) Relief of the Central American Isthmus with core locations (white dots) and
561 approximate location of low-level wind-jets (gap winds, purple arrows).
562 Climatological QuikScat wind vectors (2000/01 - 2008/12 inclusive) show wind

563 direction and speed for March (a) and September (b). QuikScat data are produced by
564 Remote Sensing Systems and sponsored by the NASA Ocean Vector Winds Science
565 Team. Data are available at www.remss.com. The lower panels show mean monthly
566 SST for March (c) and September (d) based on satellite estimates (AVHRR Pathfinder
567 Version 5.0 data; Casey et al., 2010). Numbers next to core sites in (c) show the
568 leading long-term EOF1 pattern.

569

570 **Figure 2.** Benthic $\delta^{18}\text{O}$ records (*Uvigerina peregrina*) and LR04 stack (Lisiecki and
571 Raymo, 2005) for (a) core ME43 on its original age scale (Benway et al., 2006), (b)
572 core ME43 (c) core ME24 and (d) core ME27 on age scales as updated in this study.
573 Black triangles indicate AMS-14C dates, black circles indicate benthic tie-points and
574 grey bars indicate the Los Chocoyos Ash Layer (Drexler et al., 1980).

575

576 **Figure 3.** Original UK'37-SST time series used for the deglacial EOF. Data for core
577 ME24 are from Kienast et al. (2006). Data for core JPC32-14 are from Pahnke et al.
578 (2007). Data for core MD29 are from Leduc et al. (2007). Data for cores V19-27,
579 V19-28, V19-30, V21-30 and RC11-238 are taken from Koutavas and Sachs (2008).
580 Data for cores ME27, ME43 and TR31 are from Dubois et al. (2009).

581

582 **Figure 4.** Downcore reconstruction of UK'37-SST of the three cores versus age for the
583 last 100 kyr. From top to bottom: (a) Greenland NGRIP $\delta^{18}\text{O}$ temperature record. (b)
584 ME43 UK'37-SST (original and 15-pt smoothed). (c) ME24 UK'37-SST (original and
585 15-pt smoothed). (d) ME-27 UK'37-SST (original and 15-pt smoothed). (e) The

586 principal component corresponding to the leading long-term EOF1. (f) Sea surface
587 temperature difference (Δ SST) between ME43 and ME24. (g) Δ SST between ME43
588 and ME27. (h) Δ SST between ME24 and ME27. The dashed horizontal line in (h)
589 represents the propagated analytical error above which the Δ SST is significant. The
590 color shading in (b), (c) and (d) represent the analytical error. Grey vertical intervals
591 mark the glacial stages MIS2 and 4, as indicated on the top of the figure. Dashed
592 vertical lines mark the North Atlantic ice rafting and melt-water events: Younger
593 Dryas (YD), Heinrich events (H1–H6).

594

595 **Figure 5.** Leading modes of SST variability: The upper (lower) panels show the patterns
596 (principal components) of the EOF modes of UK'37-SST reconstructions from ETP
597 cores. (a) EOF analysis based on SST reconstructions (5-100 kyr BP) from cores
598 ME24, ME27, ME43. (b) and (c) 1st and 2nd EOF analysis based on the 25-5 kyr BP
599 section of cores V19-27, V19-28, V19-30, V21-30, RC11-238, MD02-2529, JPC32-14,
600 ME24, ME43, ME27 and TR31 (Kienast et al., 2006; Leduc et al., 2007; Pahnke et al.,
601 2007; Koutavas and Sachs, 2008; Dubois et al., 2009; Dubois et al., 2011; Kienast et
602 al., 2013). The contour lines represent SST anomalies from a climate model
603 experiment conducted with the GFDL CM2.1 model (Timmermann et al., 2007),
604 representing present-day control run (100 year mean) minus the AMOC off case (100
605 year mean). (d) The principal component corresponding to the leading long-term EOF
606 (red) and the UK'37-SST reconstruction from the eastern North Atlantic core MD01-
607 2444 (Martrat et al., 2007; blue). (e) The principal component corresponding to the
608 leading deglacial EOF (red), along with equatorial insolation during October,

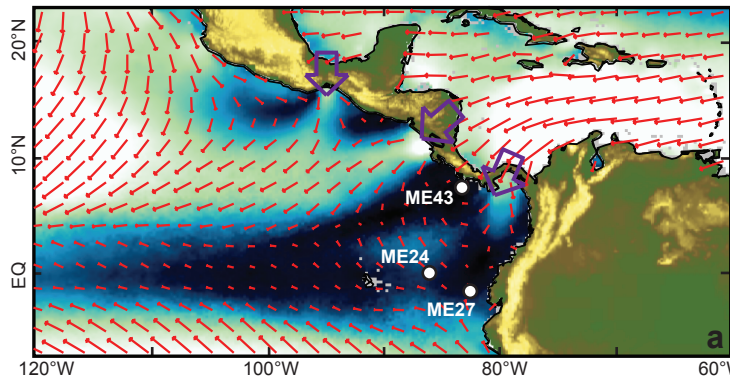
609 November, December (OND, normalized; orange dashed) and atmospheric CO₂ (Lüthi
610 et al., 2008; blue). (f) The principal component corresponding to the second deglacial
611 EOF (red), along with UK'37-SST reconstruction from eastern North Atlantic core
612 MD01-2444 (Martrat et al., 2007; blue).

613

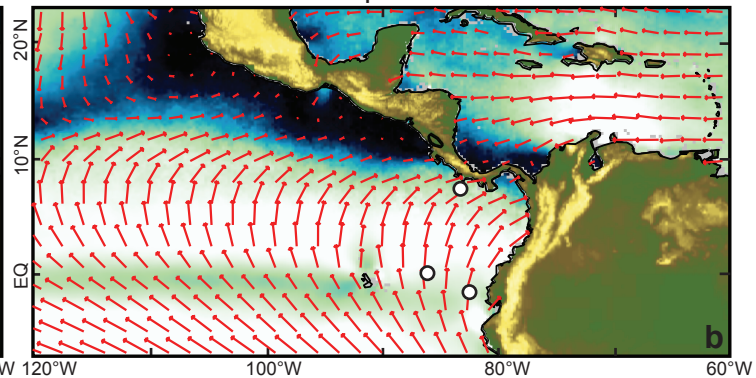
614 **Figure 6.** Comparison of the UK'37-SST from ME43 and MD29 (Leduc et al., 2007),
615 both located in the Panama Basin.

1. Millennial-scale cooling events occur in the East Pacific during Heinrich events
2. N. Atlantic SST anomalies affect Central American gap winds and thus the E. Pacific
3. Hydrographic conditions were similar during glacial and interglacial periods.

March



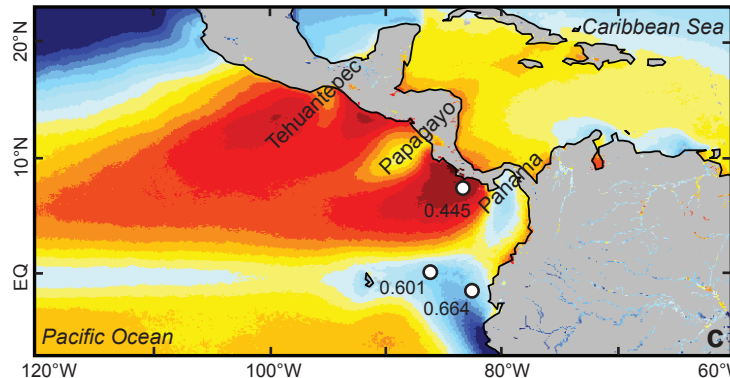
September



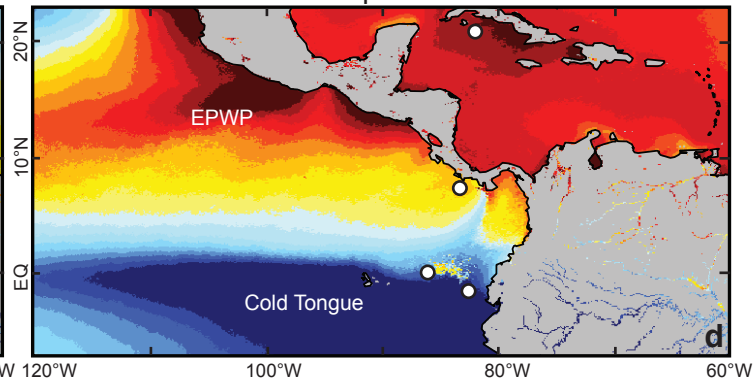
Wind speed (m/s) → 5.7



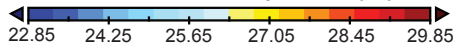
March

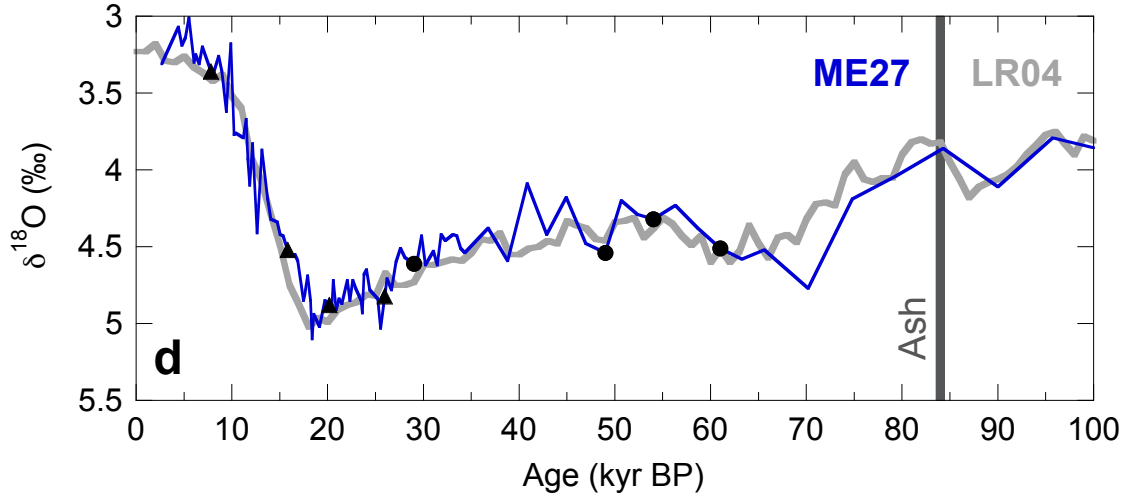
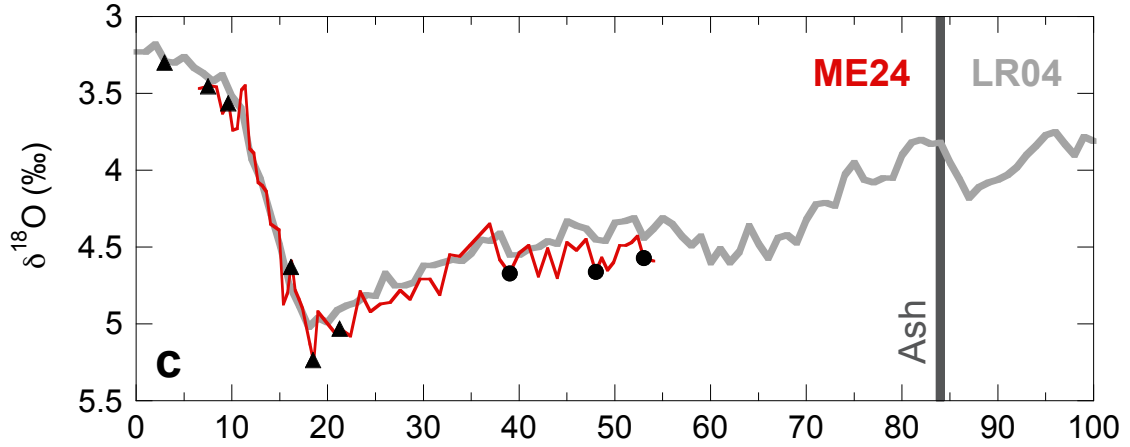
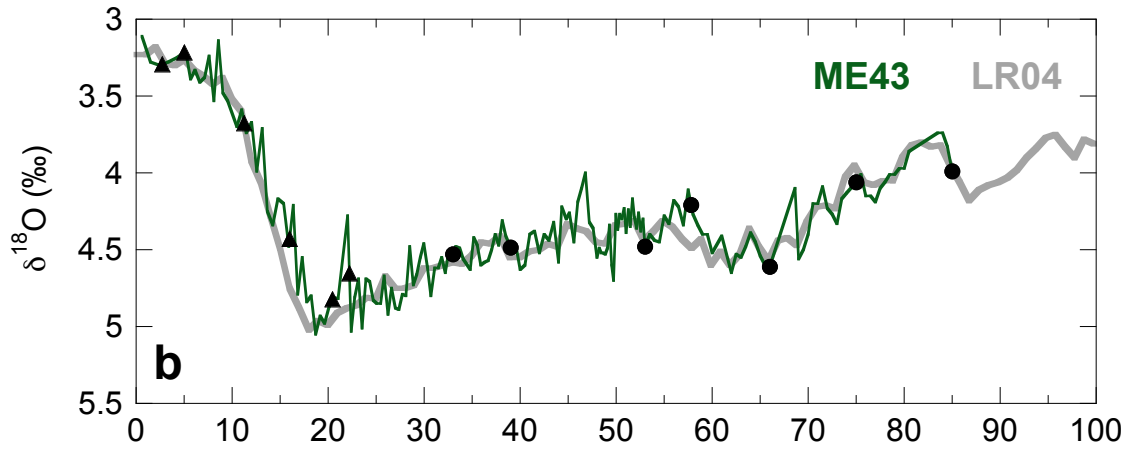
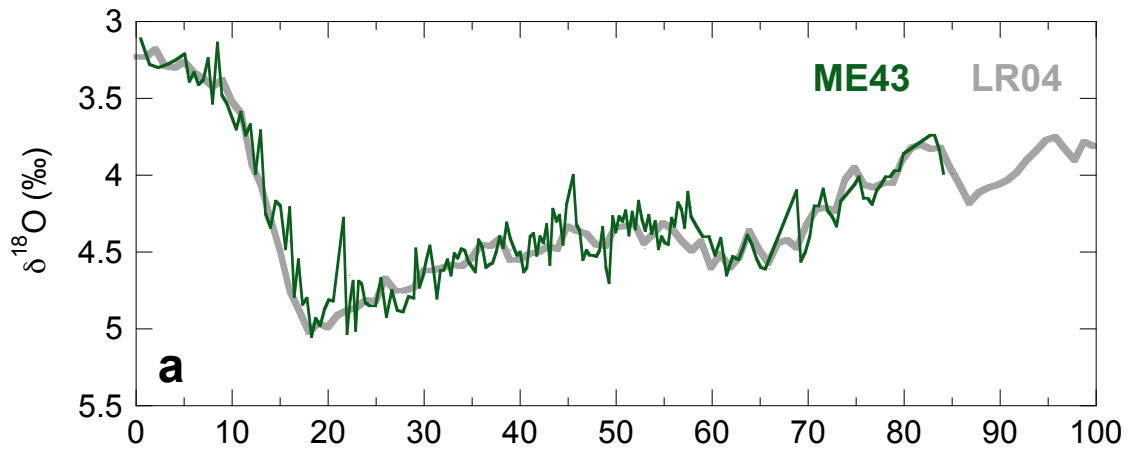


September



Sea surface temperature (°C)





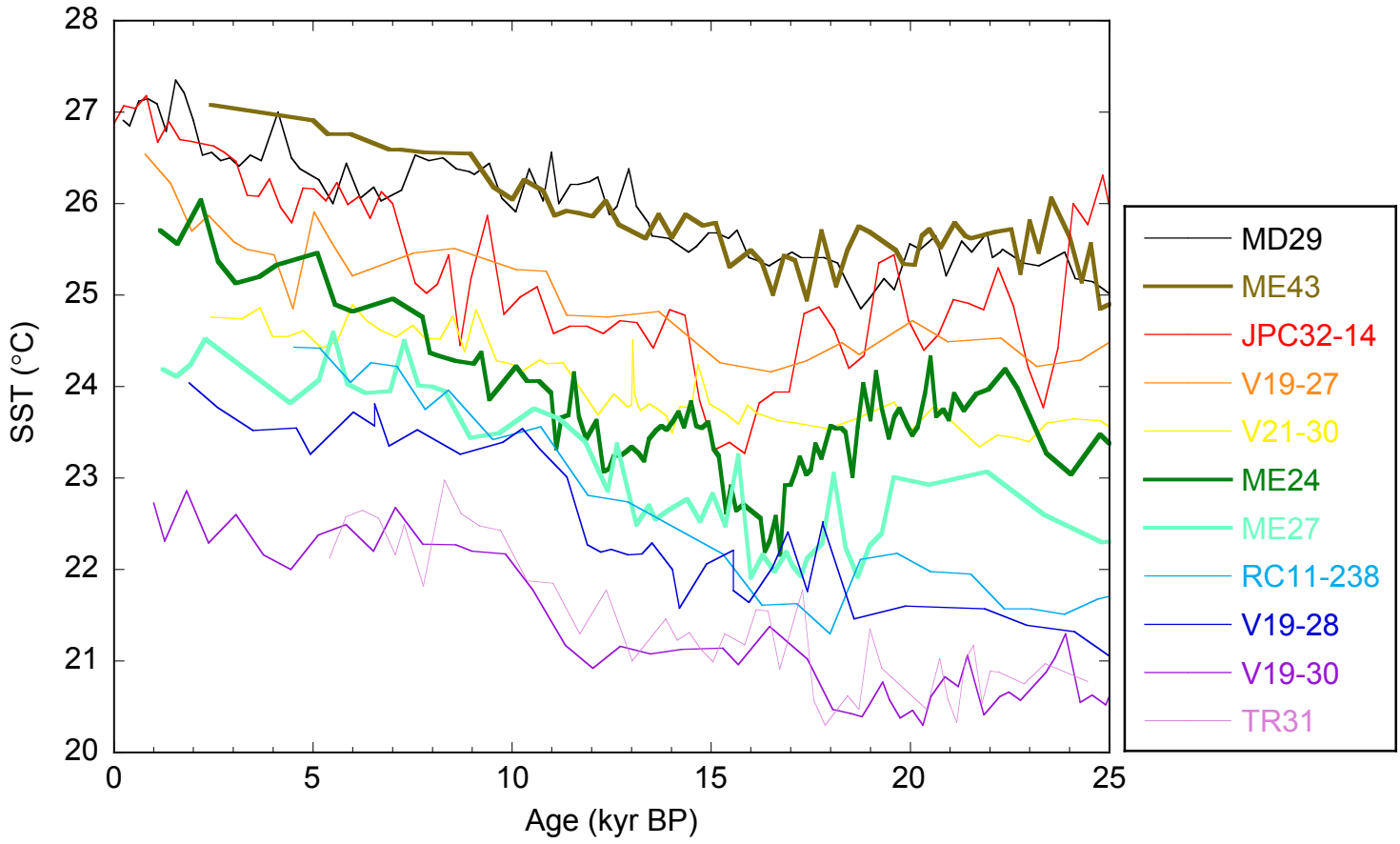
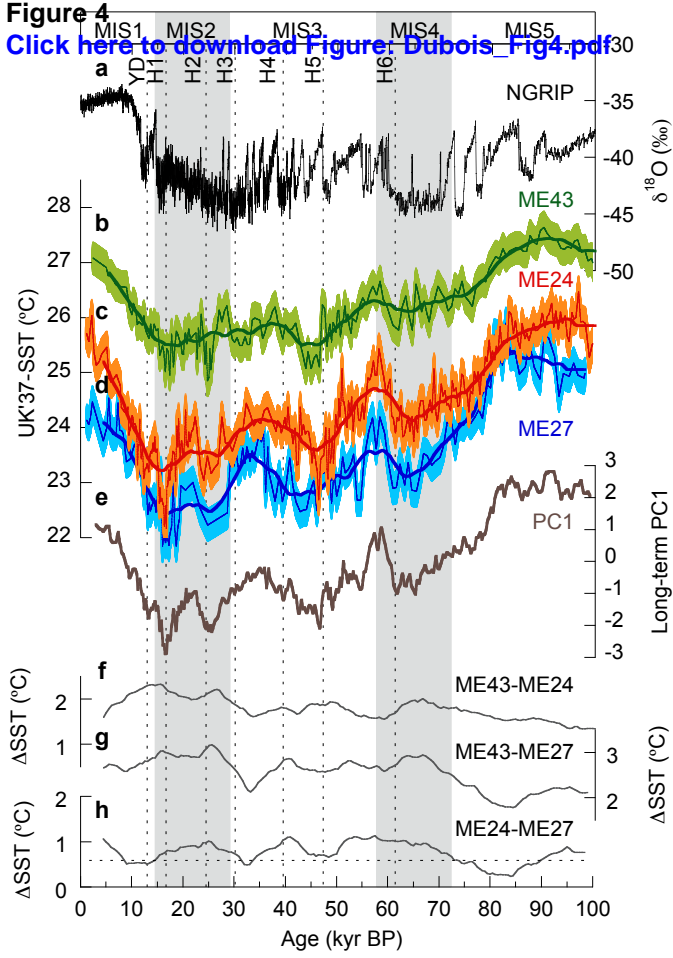
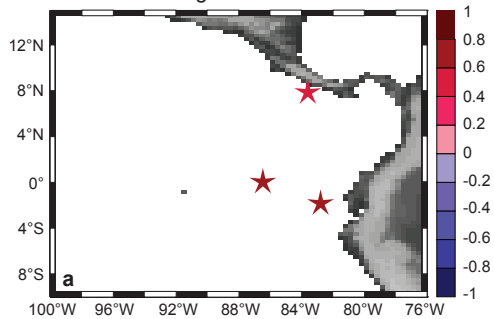
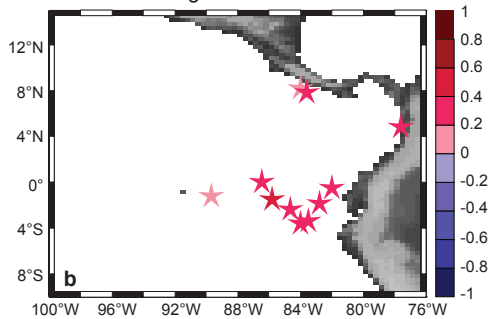


Figure 4[Click here to download Figure: Dubois_Fig4.pdf](#)

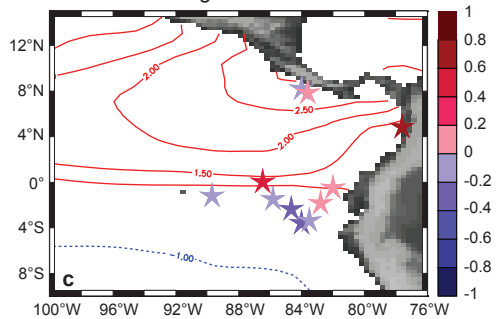
Long-term EOF1



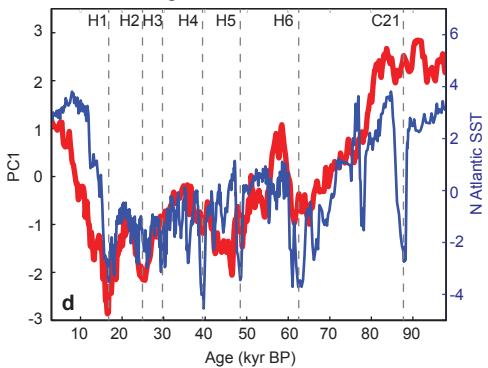
Deglacial EOF1



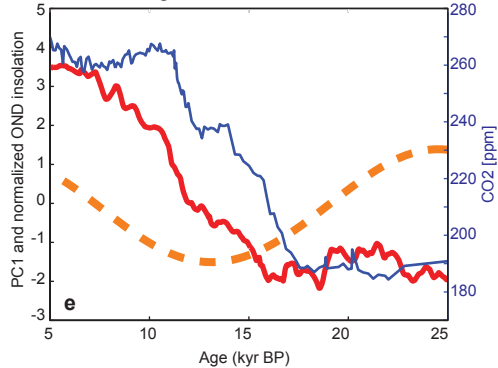
Deglacial EOF2



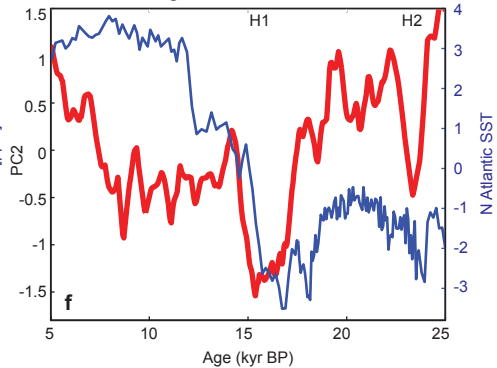
Long-term PC1 93.7%



Deglacial PC1 80.5%



Deglacial PC2 10.0%



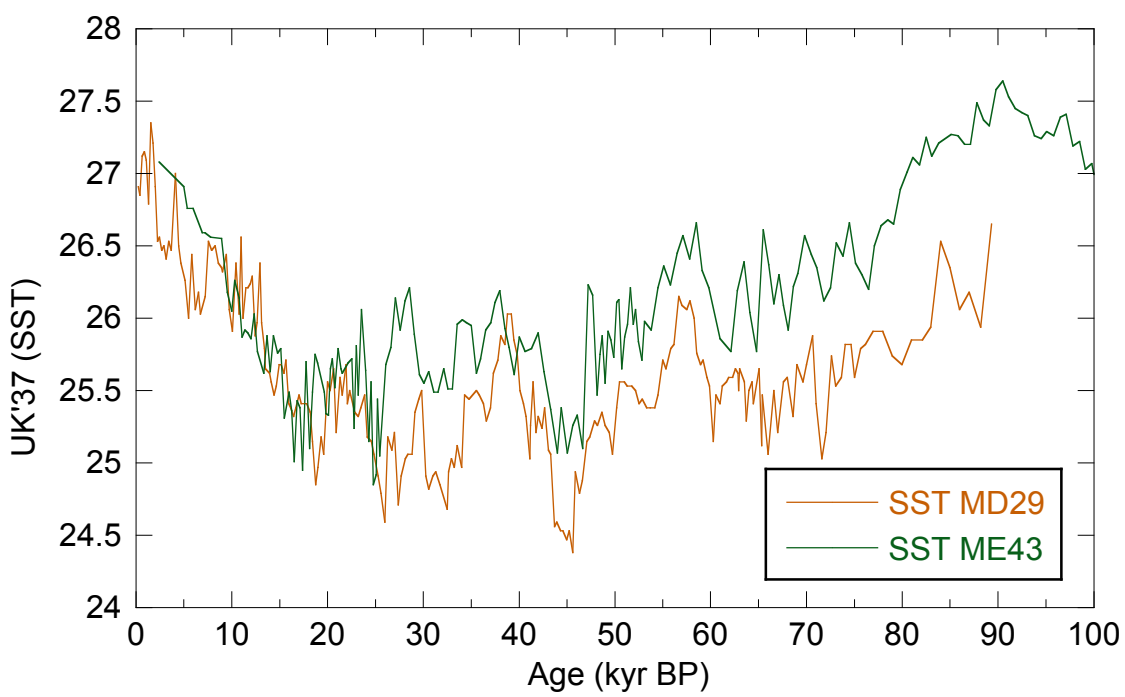


Table 1[Click here to download Table: Dubois_Table1.xlsx](#)**Table 1.** AMS-¹⁴C data

Core	Depth [cm]	¹⁴ C Age [kyr]	SD [kyr]	Calendar Age [kyr]	SD [kyr]	Species
ME-24	17.5	3.26	0.03	2.99	0.06	<i>N. dutertrei</i>
ME-24	48.5	7.10	0.04	7.52	0.04	<i>N. dutertrei</i>
ME-24	81	9.04	0.04	9.62	0.06	<i>N. dutertrei</i>
ME-24	230	13.90	0.04	16.16	0.08	<i>N. dutertrei</i>
ME-24	291	15.70	0.09	18.50	0.11	<i>N. dutertrei</i>
ME-24	351	18.05	0.07	21.24	0.14	<i>N. dutertrei</i>
ME-27	41	7.43	0.04	7.83	0.05	<i>N. dutertrei</i>
ME-27	91	13.65	0.07	15.84	0.12	<i>N. dutertrei</i>
ME-27	120	17.20	0.07	20.18	0.11	<i>N. dutertrei</i>
ME-27	160	22.20	0.10	25.97	0.09	<i>N. dutertrei</i>
ME-43	13	3.10	0.04	2.80	0.04	<i>N. dutertrei</i>
ME-43	26	4.94	0.04	5.18	0.08	<i>N. dutertrei</i>
ME-43	91	10.45	0.06	11.46	0.15	<i>N. dutertrei</i>
ME-43	131	13.70	0.08	15.90	0.13	<i>N. dutertrei</i>
ME-43	186	17.90	0.10	21.04	0.16	<i>N. dutertrei</i>
ME-43	208	19.15	0.11	22.54	0.12	<i>N. dutertrei</i>

Table 2

[Click here to download Table: Dubois_Table2.xlsx](#)**Table 2.** Age models as updated in this study

Depth (cm)	Age (kyr BP)	Dating method
ME43: Age model published by Benway et al. (2006)		
13	2.68	radiocarbon MARINE04
26	5.06	radiocarbon MARINE04
91	11.39	radiocarbon MARINE04
131	15.56	radiocarbon MARINE04
186	20.51	radiocarbon MARINE04
208	22.12	radiocarbon MARINE04
228	23.30	planktic O18 strat. (tied to GISP2)
238	24.00	planktic O18 strat. (tied to GISP2)
281	28.94	planktic O18 strat. (tied to GISP2)
286	29.14	benthic O18 strat (tied to MD95-2045)
421	38.97	benthic O18 strat (tied to MD95-2045)
516	45.53	benthic O18 strat (tied to MD95-2045)
696	57.81	benthic O18 strat (tied to MD95-2045)
736	62.74	benthic O18 strat (tied to MD95-2045)
956	83.18	benthic O18 strat (tied to MD95-2045)
ME43: Age model as updated in this study		
13	2.72	radiocarbon MARINE13
26	5.04	radiocarbon MARINE13
91	11.26	radiocarbon MARINE13
131	15.99	radiocarbon MARINE13
186	20.47	radiocarbon MARINE13
208	22.23	radiocarbon MARINE13
386	33.00	benthic O18 strat. (tied to LR04)
465	39.00	benthic O18 strat. (tied to LR04)
642	53.00	benthic O18 strat. (tied to LR04)
696	57.81	benthic O18 strat. (tied to LR04)
770	66.00	benthic O18 strat. (tied to LR04)
871	75.00	benthic O18 strat. (tied to LR04)
966	85.00	benthic O18 strat. (tied to LR04)
ME24: Age model as updated in this study		
17.5	2.99	radiocarbon MARINE13
48.5	7.52	radiocarbon MARINE13
81	9.62	radiocarbon MARINE13
230	16.16	radiocarbon MARINE13
291	18.50	radiocarbon MARINE13
351	21.24	radiocarbon MARINE13
522	39.00	benthic O18 strat. (tied to LR04)
612	48.00	benthic O18 strat. (tied to LR04)
692	53.00	benthic O18 strat. (tied to LR04)
980	84.00	Ash layer (min in CaCO ₃)
ME27: Age model as updated in this study		
41	7.83	radiocarbon MARINE13
91	15.84	radiocarbon MARINE13
120	20.18	radiocarbon MARINE13
160	25.97	radiocarbon MARINE13
173	29.00	benthic O18 strat. (tied to LR04)
271	49.00	benthic O18 strat. (tied to LR04)
301	54.00	benthic O18 strat. (tied to LR04)
331	61.00	benthic O18 strat. (tied to LR04)
431	84.00	Ash layer (min in CaCO ₃)
522	110.00	benthic O18 strat. (tied to LR04)

Table 3[Click here to download Table: Dubois_Table3.xlsx](#)**Table 3.** List of cores used in the deglacial EOF computation

Core name	Location		Depth (m)	degl. EOF1	degl. EOF2	References
	Lat.	Long.				
MD29	8°12'N	84°07'W	1619	0.1974	-0.1751	Leduc et al. (2007)
ME43	7°51'N	83°36'W	1368	0.231	0.0126	Dubois et al. (2009)
JPC32-14	4°39'N	77°57'W	2200	0.2139	0.777	Pahnke et al. (2006)
ME24	0°01'N	86°27'W	2941	0.2754	0.4135	Kienast et al. (2006)
V19-27	0°28'S	82°40'W	1373	0.2277	0.0504	Koutavas & Sachs (2008)
ME27	1°51'S	82°47'W	2203	0.3117	0.0996	Dubois et al. (2009)
V21-30	1°13'S	89°41'W	617	0.206	-0.0573	Koutavas & Sachs (2008)
RC11-238	1°31'S	85°49'W	2573	0.4787	-0.0585	Koutavas & Sachs (2008)
V19-28	2°22'S	84°39'W	2720	0.3987	-0.2706	Koutavas & Sachs (2008)
V19-30	3°23'S	83°31'W	3091	0.326	-0.1896	Koutavas & Sachs (2008)
TR31	3°37'S	83°58'W	3209	0.3167	-0.2572	Dubois et al. (2009)