

**High latitude controls on the evolution of eastern equatorial Pacific
oceanography and climate since the last glacial period**

By

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PREFACE

The eastern equatorial Pacific (EEP) amplifies changes to the global climate system through heat and carbon exchange with the atmosphere [*Hastenrath*, 1980; *Takahashi et al.*, 2002; *Philander and Fedorov*, 2003; *Boccaletti et al.*, 2004]. The physical oceanography and patterns of wind forcing across the region give rise to a uniquely strong link between cool, carbon-rich subsurface waters and warm, tropical surface waters [e.g. *Fiedler and Talley*, 2006]. Subsurface waters upwell to the surface ocean along the equator and the west coast of South America creating a tongue of cool water that stretches westward across the Pacific. The result is a net transfer of heat from the atmosphere into the ocean, elevated rates of primary production, and the release of CO₂. The EEP yields 15-50% of global open ocean productivity [*Chavez and Barber*, 1987] and releases nearly two-thirds of the marine carbon dioxide flux to the atmosphere each year [*Takahashi et al.*, 2002].

Upwelling strength and mixing between EEP subsurface and surface waters is not constant thru time. At present, temporal variability is dominated by the El Niño Southern Oscillation (ENSO), a tropical air-sea phenomenon that alters the heat distribution across the equator and the strength of upwelling in the EEP on a 2 to 7 year timescale [e.g. *Wang and Fiedler*, 2006]. The effects of ENSO are felt most strongly in the EEP, giving rise to changes in phytoplankton productivity, sea surface temperature and precipitation patterns. However, ENSO events also lead to global scale atmospheric reorganizations that affect weather patterns across the globe. In the United States, one may recall the 2015 strong El Niño event that provided a much-needed reprieve to drought-stricken California.

On decadal and longer timescales we must rely on paleoceanographic records and models to reconstruct the dynamics of the EEP because the historical record of ocean observations is short (<135 years) and sparse [Hobbs and Willis, 2013]. Thus far, proxy and model evidence indicate that remotely forced mechanisms in the southern and northern high-latitudes may overwhelm locally driven variability through changes in the transport and physio-chemical properties of subsurface intermediate waters. These water masses form at the poleward boundaries of the subtropical gyres [Liu and Philander, 1995; Gu and Philander, 1997; Lu *et al.*, 1998] and at even higher latitudes [Toggweiler *et al.*, 1991; Andreassen *et al.*, 2001]. They transport heat, salt, and nutrients from their formation region into the EEP subsurface where they drive oceanographic change from the bottom-up. This subsurface variability may then feedback to the surface ENSO-cycle in intriguing ways.

My dissertation work investigates the relative role of high- and low-latitude forcings in shaping EEP oceanography and climate since the last glacial period. The transition from the Last Glacial Maximum (LGM) to the current interglacial is a natural experiment in rapid, global warming. Global temperatures rose by approximately 3.5°C between the LGM (21 ka) and the early Holocene (6-10 ka) [Denton *et al.*, 2010; Shakun *et al.*, 2012], a rise comparable to that projected for the coming century [IPCC, 2014]. The first three chapters reconstruct change within EEP subsurface intermediate waters, while the last two focus on the surface ocean response. The most significant finding of this dissertation is that the EEP is a passive responder to subsurface change driven from the bottom-up by the high southern and northern latitudes on millennial and longer timescales. The work therefore refutes ideas that locally forced ocean-atmosphere

dynamics in the EEP drove global climate shifts over the past 30 kyrs [e.g. Clement et al., 1999].

Chapter one examines the evolution of subsurface intermediate waters in the EEP upper water column since the last glacial period. Sediment cores from three intermediate depths in the EEP are used to reconstruct change in stratification over the top 1000 m of the water column. $\delta^{18}\text{O}$ signatures measured on the carbonate shells of benthic foraminifera, zooplankton that live at the sediment-water interface, provide a combined measure of water temperature and salinity and therefore density at the depth of core recovery. Water density change arose first at depth, decreasing in phase with rising temperatures over Antarctica [Jouzel et al., 2007], and propagated up to the surface over the course of deglaciation. Early warming in the subsurface drove a reduction in stratification that increased mixing over the top 1000 m of the water column and CO_2 degassing during the last deglaciation. These findings are significant because they suggest stratification and mixing between intermediate and surface waters in the EEP is regulated by climate in the southern high latitudes.

Chapter two also focuses on the role of the subsurface in driving change in the EEP but on much shorter timescales (<200 years). This work was motivated by recent observations that expose unprecedented rates of warming in the deep ocean that track observed surface warming [Balmaseda et al., 2013]. We test a new tool, the $\delta^{18}\text{O}$ signature of individual benthic foraminifera, to evaluate the natural variability of the deep ocean. A first use of the proxy at 1000 m water depth in the EEP reveals temperature variability of more than 2°C within 200 years, significantly larger than the modern range.

Rates of modern deep water variability in the EEP are therefore not unprecedented in the Holocene.

Chapter three investigates the role of Pacific deep and intermediate waters in glacial-interglacial atmospheric CO₂ variations. The leading hypothesis suggests carbon is stored in the deep Southern Ocean during glacial periods and is released back to the atmosphere via Southern Ocean Intermediate Waters (SOIW) [Toggweiler *et al.*, 2006; Marchitto *et al.*, 2007; Anderson *et al.*, 2009]. Initial tests of the hypothesis at intermediate depths in the Pacific were affirmative but recent proxy records and model runs have cast doubts [Hain *et al.*, 2011; De Pol-Holz *et al.*, 2010; Rose and Sikes, 2010]. In chapter three I propose a new mechanism for carbon release during the last deglaciation that reconciles all but one [Stott *et al.*, 2009] previously published radiocarbon record under a single unified theory.

The relative impact of remotely forced subsurface change versus local air-sea dynamics on surface ocean processes in the EEP is explored in Chapters four and five. We reconstruct sea surface temperature, thermocline depth, and primary productivity at four sites spread across the EEP. In Chapter four the relationships between these parameters are evaluated, for the first time in the paleo-record, to determine whether the mean state of the EEP can be regarded as “scaled-up” El Niño or La Niña, or a state similar to neither. Correlations between parameters indicate that ENSO was not the dominant source of variability in surface ocean properties on millennial timescales over the past 30 kyrs.

Chapter 5 focuses on the response of phytoplankton in the EEP to global change over the past 30 kyrs. Productivity records from previous studies exhibit a regionally

heterogeneous and at times conflicting response to global climate change since the Last Glacial Maximum, which makes it difficult to assess the role of the biological pump in atmospheric CO₂ change. Carbon balance in the EEP is determined by the export ratio of organic carbon to calcium carbonate, which can be estimated as the relative abundance of diatoms (non-calcareous) versus coccolithophores (calcareous) [e.g. *Kohfeld et al.*, 2005]. Here, the productivity response of diatoms and coccolithophores is reconstructed at multiple sites in the EEP to show that diverging production histories are not caused by differing responses of diatoms and coccolithophores but reflect true spatial heterogeneity related to the proximity of local iron sources. Thus, the EEP biological pump was not an important carbon sink during the last glacial period and cannot be relied upon to mitigate anthropogenic climate change in the future.

The work presented in this dissertation demonstrates a primary role for high-latitude climate in shaping the EEP since the last glacial period. Its fingerprint is observed at intermediate depths and the surface ocean using a variety of paleoclimate proxies. This finding is important for predicting the response of the EEP to future climate change. Based on the results presented in this dissertation, I expect warm subsurface anomalies from the high latitudes will drive a reduction in upper ocean stratification in the EEP and a mean deepening of the equatorial Pacific thermocline. These changes in upper ocean structure will subsequently lead to weakened upwelling of cool subsurface waters and reduced heat uptake in the EEP over the coming centuries.

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List of Abbreviations

Currents

Ecuador Peru Countercurrent	EPCC
Equatorial Undercurrent	EUC
North Equatorial Countercurrent	NECC
Peru Coastal Current	PCC
Peru-Chile Undercurrent	PCUC
South Equatorial Current	SEC

Oceanographic Zones

Eastern Pacific Warm Pool	EPWP
Equatorial Upwelling	EU
Equatorial Cold Tongue	ECT
Galapagos Upwelling	GU
Peru Coastal Upwelling	PCU

Water Masses

Antarctic Intermediate Water	AAIW
Antarctic Surface Waters	AASW
Equatorial Pacific Intermediate Water	EqPIW
Equatorial Surface Water	ESW
North Pacific Deep Water	NPDW
North Pacific Intermediate Water	NPIW
Pacific Deep Water	PDW
South Pacific Deep Water	SPDW
Southern Ocean Intermediate Water	SOIW
Subantarctic Mode Water	SAMW
Subtropical Mode Water	STMW

Time Intervals

Antarctic Cold Reversal	ACR
Bolling Allerod	B/A
Heinrich Stadial 1	HS1
Last Glacial Maximum	LGM
Younger Dryas	YD

Miscellaneous

Eastern Equatorial Pacific	EEP
El Nino Southern Oscillation	ENSO
Gas Chromatograph Flame Ionization Detector	GC-FID
Oxygen Minimum Zone	OMZ
Sea Surface Temperature	SST

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CHAPTER ONE

Links between Eastern Equatorial Pacific stratification and atmospheric CO₂ rise during the last deglaciation

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Abstract

It is difficult to untangle the mixed influences of high and low latitude climate forcing in the eastern equatorial Pacific (EEP). Here we test the hypothesis that the Southern Ocean drove change in the EEP via subsurface intermediate waters during the last deglaciation. We use the $\delta^{18}\text{O}$ signature of benthic foraminifera to reconstruct water density changes during the last 25 kyrs at three intermediate water depths (370 m, 600 m, and 1000 m) in the EEP. Carbonate $\delta^{18}\text{O}$ records a combined signature of temperature ($0.25\text{‰ } \delta^{18}\text{O}_{\text{shell}} / ^\circ\text{C}$) and salinity ($0.24\text{‰ } \delta^{18}\text{O}_{\text{shell}} / \text{psu}$) and is therefore more closely related to density than temperature or salinity alone. We find that benthic foraminiferal $\delta^{18}\text{O}$ values decreased first in the subsurface, simultaneously with rising temperatures over Antarctica, and propagated up to the surface within ~ 3 kyrs. The early subsurface response initiated a rapid decrease in density stratification over the upper water column as indicated by reduced $\delta^{18}\text{O}$ gradients between surface and intermediate depths. Stratification of the upper water column remained low through the termination, with stratification minima reached during Heinrich Stadial 1 and the Younger Dryas (YD), synchronous with the two-part deglacial rise in atmospheric CO_2 . Centennial-scale shifts towards heavier $\delta^{18}\text{O}$ signatures at 370 and 600 m during the YD indicate short-lived shifts in the Subantarctic Mode Water/Antarctic Intermediate Water boundary to shallower intermediate depths. We suggest decreased density gradients during the deglaciation accelerated vertical mixing across the EEP, and potentially the entire South Pacific subtropical gyre, which enhanced CO_2 delivery from depth to the surface ocean and atmosphere.

1. Introduction

The last deglaciation represents earth's most recent natural experiment in rapid global climate change [e.g. *Denton et al.*, 2010]. Beginning at 18 ka, a relatively minor change in the amount of incoming solar radiation to the high northern latitudes triggered a series of climate feedbacks (such as ice-albedo and CO₂) that amplified and distributed high latitude temperature anomalies through the climate system, resulting in a 4°C rise in global mean surface temperatures, large-scale changes in global patterns of ocean circulation, and a two-step rise in atmospheric CO₂ [*Imbrie et al.*, 1993; *McManus et al.*, 2004; *Denton et al.*, 2010; *Jouzel et al.*, 2007; *Lourantau et al.*, 2010; *Shakun et al.*, 2012; *Marcott et al.*, 2014]. Ice cores recovered from Greenland and Antarctica provide well-constrained records of the timing and nature of these transitions [*Grootes et al.*, 1993; *Jouzel et al.*, 2007; *Lourantau et al.*, 2010; *Marcott et al.*, 2014]; however, we still have few constraints on how high latitude change propagated spatially over time. One key unknown is how and to what extent high latitude forcing influenced low latitude climate, particularly in regions where surface and shallow subsurface properties have strong links to global climate, such as the equatorial Pacific [*Clement and Cane*, 1999; *Rosenthal et al.*, 2003].

Dynamical shifts in the behavior of the EEP amplify changes to the global climate system through their control on heat and carbon exchange with the atmosphere.

Upwelling from beneath a strong shallow thermocline regulates the exchange of cool, nutrient-rich intermediate waters with the surface ocean, exerting a first order control on the regional air-sea CO₂ flux as well as the efficiency of Pacific meridional heat transport [*Hastenrath*, 1980; *Takahashi et al.*, 2002 *Philander and Fedorov*, 2003; *Boccaletti et al.*,

2004]. Heat and carbon exchange with the atmosphere is regulated by the efficiency of exchange between intermediate and surface waters and therefore depends strongly on oceanic circulation and stratification patterns. Subsurface variability, arising at the poleward boundary of the subtropical gyres [*Liu and Philander, 1995; Gu and Philander, 1997; Lu et al., 1998*] and even higher latitudes via intermediate waters [*Toggweiler et al., 1991; Andreasen et al., 2001*] may play an essential role in setting the background stratification state of the EEP.

Both high and low latitude climate forcings are important to EEP dynamics, which makes the relative role of tropical versus high latitude influences and high latitude southern versus high latitude northern hemisphere influences difficult to parse during the last deglaciation. On short timescales, locally forced ocean-atmosphere interactions regulate the strength and direction of the prevailing winds and, in turn, patterns of upwelling in the region. There is a strong seasonal cycle in the EEP and the El Niño Southern Oscillation represents the primary source of interannual variability [*Philander, 1985; Clement and Cane, 1999; Lavin et al., 2006*]. On centennial to millennial timescales, however, remotely forced mechanisms may overwhelm locally driven variability. These mechanisms include: (1) changing interhemispheric temperature gradients that are capable of driving long-lived N-S shifts in the mean position of the easterlies and thus the zones of upwelling within the EEP via shifts in the Hadley circulation (i.e. the atmospheric bridge) [*Pahnke et al., 2007; Koutavas and Sachs, 2008*] and (2) changes in the transport and/or physio-chemical properties of subsurface waters via oceanic tunnels [*Liu and Yang, 2003*]. The efficiency and relative importance of these two pathways are currently not well-constrained.

Using a suite of well-dated, sediment cores from three intermediate water depths, we evaluate the response of the EEP upper water column to deglacial climate change to assess the role of the oceanic tunnel in driving oceanographic change in the EEP. Benthic foraminiferal $\delta^{18}\text{O}$ records provide a combined signal of temperature and salinity, and therefore density at the surface, 370 m, 600 m, and 1000 m water depth in the equatorial cold tongue region throughout the last 25 kyrs. An alkenone record estimates sea surface temperature (SST) for the last 25 kyrs. Together, these records provide direct evidence for significant changes in EEP intermediate water structure and stratification during the last deglaciation at high temporal resolution with important implications for regional heat and carbon balance. We diagnose high versus low latitude driven change by comparing the timing and relative magnitude of warming across the EEP upper water column; locally forced coupled ocean-atmosphere interactions will produce the largest recorded signal of oceanographic change at the sea surface (i.e. top-down forcing), while remotely forced oceanic mechanisms are observed first, and most strongly in the subsurface (i.e. bottom-up forcing). In the EEP, we find the largest and earliest signals of deglacial warming in the subsurface, which provides support for the role of the Southern Ocean in driving deglacial change in the EEP water column from the bottom up.

2. Modern Oceanographic Setting

The EEP is a region of complex currents forced primarily by the local wind field and intermediate water advection [*Wyrski*, 1966; *Fiedler and Talley*, 2006; *Kessler*, 2006]. The southeast trade winds drive surface currents and Ekman divergence along the equator and the South American coast [*Wyrski*, 1966; *Kessler*, 2006]. Upwelling velocities along the Peru Margin are strongest and maintain the equatorial cold tongue, a

region of anomalously cold sea surface temperatures just south of the equator [Kessler, 2006]. In the subsurface, the eastward flowing Equatorial Undercurrent (EUC), centered at ~80 m depth, flows from west to east across the Pacific Basin transporting modified Subtropical Mode Water (STMW). Below the EUC, lies the thermostad, a region of relatively uniform temperature (11-14°C) between ~150 and 300 m depth in the EEP [Tsuchiya, 1981]. More than 80% of the thermostad water is derived from mixing between waters above (STMW) and below the thermostad (Subantarctic Mode Water/Equatorial Pacific Intermediate Water), with horizontal advection contributing much less by volume (Qu *et al.*, 2009). Some evidence suggests thermostad waters do reach the surface mixed layer in the EEP [Qu *et al.*, 2009]; however, the influence of thermostad waters on regional SSTs appears negligible during at least the late Holocene [Kalansky *et al.*, 2015]

Subantarctic Mode Water (SAMW) and Equatorial Pacific Intermediate Water (EqPIW), a mixture of AAIW and Pacific Deep Water, dominate at water depths between 300 and 1200 m in the EEP [Sarmiento *et al.* 2004; Fiedler and Talley, 2006]. SAMW overlies EqPIW between about 300 and 600 m in the EEP where it experiences strong vertical mixing at the base of the thermostad [Fiedler and Talley, 2006; Kalansky *et al.*, 2015]. SAMW and AAIW (a major component of EqPIW) are sourced from the southern high latitudes and transported into the EEP through an oceanic tunneling system (Figure 1) [Toggweiler *et al.*, 1991; Bostock *et al.*, 2010; 2013]. SAMW forms in the Subantarctic Zone, within the northernmost Antarctic Circumpolar current during late winter convective overturning and travels from the Southern Ocean northwestward to join the New Guinea Coastal Undercurrent, eventually flowing eastward across the equatorial

Pacific into the EEP subsurface (Figure 1) [McCartney, 1975; Toggweiler *et al.*, 1991; Rintoul and England, 2002; Qu *et al.*, 2009; Bostock *et al.*, 2010; 2013]. Today, SAMW temperature and salinity signatures are not observed north of 30°S (Figure 1) [Herraiz-Borreguero and Rintoul, 2011]. However, water column radiocarbon ($\Delta^{14}\text{C}$) data from the Peru margin and neodymium and $\delta^{13}\text{C}$ reconstructions from the EEP indicate that SAMW penetrates to the equator today and has done so for at least the last 30 kyrs [Spero and Lea, 2002; Toggweiler *et al.*, 1991; Pena *et al.*, 2013]. These data indicate that sufficiently strong SAMW temperature and/or salinity signals contribute to equatorial subsurface ocean heat content and stratification anomalies [Kalansky *et al.*, 2015].

EqPIW dominates at intermediate depths in the EEP below ~600 m. Although AAIW is the main component of EqPIW, the geochemical signature of EqPIW is clearly distinct from that of AAIW. EqPIW has a higher minimum salinity (34.5-34.6 psu), higher nutrients, higher silicate, higher DIC, lower oxygen, and an older $\Delta^{14}\text{C}$ signature, which indicate mixing between AAIW and Pacific Deep Water in the EEP [Bostock *et al.*, 2010; 2013]. Despite this, AAIW is still a major conveyor of heat and salt to the intermediate depth EEP, bringing temperature and salinity signals from the Southern Ocean where the water mass forms in three hotspots: (1) southwest and east of New Zealand, (2) west of the East Pacific Rise, and (3) west of the Drake Passage (Figure 1) [Bostock *et al.*, 2013; Herraiz-Borreguero *et al.*, 2011]. The main AAIW formation region is the southeast Pacific site west of the Drake Passage where the coldest and densest forms of SAMW mix with Antarctic Surface Waters (AASW) south of the

subantarctic front (Figure 1) [McCartney, 1975; Sloyan and Rintoul, 2001; Bostock et al., 2010; 2013; Hartin et al.; 2011].

3. Site Locations

A series of sediment cores were collected during a cruise of the R/V Knorr 195-5 to the eastern equatorial Pacific cold tongue in 2009. We analyzed four cores representing three water depths to evaluate changes in the EEP subsurface structure through time. CDH 23 (03° 44.95 S, 81° 08.05 W) and CDH 26 (03°59.16 S, 81°18.52 W) were recovered along the Peru Margin at 373 and 1023 m depth, respectively (Figure 2). Cores CDH 41 (01° 15.94 S, 89° 41.88 W) and GGC 43 (01°15.13 S, 89° 41.07), were recovered at 595 and 617 m water depth from the Galapagos Platform, just north of Española Island. At present, the CDH 23 core depth (370 m) lies within SAMW, CDH 41/GGC 43 recovery depths lie near the boundary between SAMW and EqPIW, and CDH 26's coring site is bathed by EqPIW (Figure 2).

4. Methods

4.1 Age Models

Age models were constructed by applying a polynomial fit to a set of planktonic radiocarbon dates for each core (Figure 3). All AMS ^{14}C dates were analyzed at the NOSAMS facility at Woods Hole Oceanographic Institute. Each radiocarbon date represents an average of 200-250 individuals, picked from the $>150\mu\text{m}$ size fraction. The age model for the Galapagos composite (GGC 43 and CDH41) is based on 26 of 33 radiocarbon measurements made on the planktonic foraminifer *Globigerinoides ruber*. The cores are spliced together at 20.2 ka, which corresponds to 365 cm in core GGC 43 and 95.25 cm in core CDH 41. Seven radiocarbon dates were excluded from our

calculation of the CDH 41 age model. Two of these dates, at 215.25 and 235.25 cm, were thousands of years too young (Figure 3b) and require movement of younger foraminifera through more than 150 cm of sediment. Given the lack of obvious bioturbation features in the core, these anomalously young ^{14}C dates may arise due to deformation of the sediments during coring. We applied a polynomial fit to the remaining ^{14}C dates and calculated the average offset between the predicted age and the ^{14}C calibrated calendar age at each age control point. Five dates, whose offset from the fit exceeded the average plus the standard deviation of the offsets (>1150 years), were removed from the final age model (Figure 3b). The age profiles for cores CDH 23 and CDH 26 are constrained with 24 and 25 AMS ^{14}C measurements, respectively, made on the thermocline dwelling foraminifer, *Neogloboquadrina dutertrei*. Planktonic foraminifera are scarce in the Peru Margin cores with only *N. dutertrei* consistently abundant. AMS ^{14}C ages were converted to calendar ages using the calibration curve of *Fairbanks et al.* [2005]. A constant reservoir age of 500 years is assumed for all measurements. Higher order polynomial fits were tested but only assigned when the R^2 values were statistically different from that of the lower order using an F test.

The age model for CDH 26 was further tuned to a nearby core, CDH 23, using the Match 2.3.1 software package to create a composite and internally consistent chronology for the two cores [*Lisiecki and Lisiecki*, 2002]. Our match is based on alkenone sea surface temperature and nitrogen isotope records from each core and thus explicitly assumes synchronicity between the two sites. This assumption is reasonable because CDH 23 and CDH 26 are located just 32 km apart. Based on our composite fit, we did not

identify any outliers among the CDH 26 AMS ^{14}C dates and therefore revert to the original polynomial fit for age control.

Our age models indicate higher sedimentation rates along the Peru Margin relative to the Galapagos Platform. The Galapagos cores, GGC 43 and CDH 41 have average sediment accumulation rates of circa 20 cm/kyr and 14 cm/kyr, respectively, while the Peru Margin cores each have sedimentation rates well over 100 cm/kyr. CDH23, the shallower core from the Peru Margin, exhibits the highest sedimentation rate, approaching 130 cm/kyr, which is presumably due to sediment trapping at shallower depths due to late deglacial and Holocene sea level rise. CDH 26 has a slightly lower average sedimentation rate (~120 cm/kyr) than the shallower core but its record extends to 25 ka.

4.2 Alkenone SST estimates

A ~25 kyr high-resolution alkenone-inferred sea surface temperature record was produced from CDH 26 and CDH 23. Samples were taken every 4 cm from the top 2.5 m of CDH 26 and the top 1.7 m of CDH 23, providing a measurement approximately every 35 years. One to two grams of freeze-dried sediment were extracted in a Dionex 200 Accelerated Solvent Extractor and analyzed on a Gas Chromatograph Flame Ionization Detector, using a modified method of *Herbert et al.* [1998]. Temperatures were calculated using the $U^{K'}_{37}$ index and the calibration of Müller et al. [1998]. Lab analytical error is equivalent to $\pm 0.1^\circ\text{C}$ based on 55 replicate analyses of a laboratory standard. Approximately 10% of our analyzed samples were run in duplicate with an average $U^{K'}_{37}$ -based temperature reproducibility of $\pm 0.05^\circ\text{C}$.

4.3 Benthic Stable Isotope Records

Benthic foraminiferal stable isotope measurements were taken at ~350-year time steps in cores CDH 26, CDH 41, and GGC 43. The higher sedimentation rate of CDH 23 allowed higher resolution, with measurements taken approximately every 120 years. Samples were freeze-dried, re-soaked in DI water, and wet-sieved through a series of mesh sieves to separate size fractions. Three to four specimens of the benthic foraminifer *Uvigerina peregrina* were picked from the 212-355 μm size fraction for all cores. Samples were analyzed at Brown University on a Finnigan MAT 252 isotope ratio mass spectrometer with a Kiel III carbonate device. Each sample was measured in duplicate, with an average standard deviation of $\pm 0.15\text{‰}$ for $\delta^{18}\text{O}$ and $\pm 0.13\text{‰}$ for $\delta^{13}\text{C}$ measurements. $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values are expressed relative to the Vienna Pee Dee Belemnite standard. The isotopic time series discussed below reflects the average of the replicates. Three outliers were removed from the CDH 23 time series due to small sample size.

4.4 Benthic Temperature Records

Benthic foraminiferal oxygen isotope compositions reflect both seawater $\delta^{18}\text{O}$ (salinity and ice volume, $\sim 0.24\text{‰}$ $\delta^{18}\text{O}_{\text{shell}}$ /psu) [Fairbanks *et al.*, 1982; Conroy *et al.*, 2014] and the ambient temperature of formation (0.25‰ $\delta^{18}\text{O}_{\text{shell}}$ / $^{\circ}\text{C}$), providing a record of seawater density at the depth of recovery [Lynch-Stieglitz *et al.*, 1999a; 1999b]. We approximate the temperature component by estimating the $\delta^{18}\text{O}_{\text{sw}}$ evolution using modern $\delta^{18}\text{O}_{\text{sw}}$ profiles acquired from bottle samples recovered via CTD (Conductivity, Temperature and Depth) casts on station during coring and correcting these values for ice volume using the Barbados sea level record [Peltier and Fairbanks, 2006]. Bottle

samples were analyzed for water isotopes at Brown University on a Picarro L1102-i isotopic water liquid analyzer.

This treatment of the data explicitly assumes no local hydrologic salinity imprint on the $\delta^{18}\text{O}$ values through time. This is a reasonable interpretation because (1) modern vertical and lateral salinity gradients are much weaker than corresponding temperature gradients [WOA13] and (2) records from the Southern Ocean indicate little influence of local hydrologic conditions on subantarctic seawater $\delta^{18}\text{O}$ records [Mashiotto *et al.*, 1999]. Glacial-interglacial temperature change accounts for 40-60% of the $\delta^{18}\text{O}$ signal in subantarctic planktonic foraminifera, with the remainder dominated by the global ice volume changes [Mashiotto *et al.*, 1999]. We therefore converted benthic stable isotope values to temperature ($^{\circ}\text{C}$) using the paleotemperature equation of Bemis *et al.* [2002] calibrated specifically for *Uvigerina spp.* in the Santa Barbara Basin.

$$T (^{\circ}\text{C}) = 21.6 - 5.50 \times (\delta^{18}\text{O}_c - \delta^{18}\text{O}_{sw})$$

The calibration range for this relationship, 3 to 8°C , is appropriate for the temperature range observed at intermediate depths in the EEP [Bemis *et al.*, 2002]. Duplicate *Uvigerina* ($\delta^{18}\text{O}_c$) values were averaged and a constant offset of -0.3‰ was applied to bring core top values in line with modern values. Corrected core top temperature estimates based on benthic $\delta^{18}\text{O}_{sw}$ agree within $\pm 0.8^{\circ}\text{C}$ of directly measured modern values. Alternate calibrations were also applied [Shackleton, 1974; Marchitto *et al.*, 2014] but ultimately rejected because calculated temperatures were unrealistically low.

5. Results

5.1 Stable Isotope Records

5.1.1 Oxygen isotopes

Over the past 25 kyrs, benthic foraminiferal $\delta^{18}\text{O}$ values decreased towards the present day at all water depths in the EEP; however, the timing and nature of the transition to interglacial values differs between records (Figure 4a). We determine the timing of the transition in each record using the Change-Point Analyzer software package, which uses mean square error estimates and bootstrapping to detect change [Taylor, 2000]. For each change point, we provide the most likely value as well as the 95% confidence interval in the main text.

The record from 1000 m water depth exhibits the heaviest benthic $\delta^{18}\text{O}$ values as well as the largest range of values (1.7‰). Glacial $\delta^{18}\text{O}$ values reached a maximum at 21 ka before a rapid but steady shift to lighter values began at 17.6 ka (17.6-17.7, $p < 0.05$). The transition from 17.6 to 14.9 ka accounts for nearly half of the total decrease in $\delta^{18}\text{O}$ over the full record. Values stabilized from 14.9 to 13.6 ka, before decreasing towards present day. Minimum $\delta^{18}\text{O}$ values are reached during the early and mid-Holocene together with a decrease in short-term variability.

Shallower in the water column, at 600 m, benthic $\delta^{18}\text{O}$ values began to decrease at 16.6 ka (16.4-17.0 ka, $p < 0.05$), approximately 1000 years later than at 1000 m. The early deglacial period, (16.6 to 14.3 ka) produces just over a third of the total $\delta^{18}\text{O}$ decrease. From 14.3 ka to 12.3 ka we observe centennial-scale oscillations, up to $0.3 \pm 0.15\text{‰}$. This time period is also characterized by a minimum offset in $\delta^{18}\text{O}$ values between 600 and 1000 m. At the end of this period $\delta^{18}\text{O}$ values shift rapidly, decreasing by $0.5 \pm 0.15\text{‰}$ in

just 500 years. This 500-year shift accounts for nearly half of the total $\delta^{18}\text{O}$ decrease during the deglaciation. After, $\delta^{18}\text{O}$ gradually decreases throughout the Holocene portion of the record.

The benthic $\delta^{18}\text{O}$ record from 370 m does not extend to the LGM, but does cover the late deglacial and Holocene time periods in great detail. As seen at 600 m, centennial-scale variability is observed from 13.9-12.3 ka at 370 m. Then, after a rapid 300 yr transition to lower values beginning 10.8 ka, $\delta^{18}\text{O}$ values decreased towards the middle Holocene, reaching a minimum at 4.8 ka.

5.1.2 Carbon Isotopes

The carbon isotope records from the Peru Margin show much larger variability over the last glacial-interglacial cycle than the records from the Galapagos platform (Figure 4b). The Peru Margin cores exhibit an increasing trend towards present day with centennial-scale fluctuations up to 0.72‰ throughout the record. In contrast, the Galapagos cores vary by just 0.35‰ over the entire 25 kyr record. At 1000 m water depth we observe two periods of prolonged $\delta^{13}\text{C}$ depletion: (1) Late Glacial to Early Deglacial (~22 -16 ka) and (2) Late Deglacial (~14.2-12.7 ka). The record from 370 m suggests $\delta^{13}\text{C}$ values continued to increase through the late Holocene until 8 ka when $\delta^{13}\text{C}$ values stabilized at an average value of $0.13 \pm 0.13\text{‰}$. $\delta^{13}\text{C}$ values at 370 m are consistently lighter than those observed at 600 and 1000 m, but records from 600 and 1000 m do not follow the expected heavy-light trend from the surface to depth. This is likely related to change in the strength and extent of the Peru Margin oxygen minimum, with the biggest impacts of metabolically-derived dissolved inorganic carbon (DIC) felt closest to the Peru Margin.

5.2 Temperature Records

5.2.1 Alkenone Sea Surface Temperatures

Our alkenone SST record from the Peru Margin exhibits a 3.3°C increase over the last 25 kyrs (Figure 5). Our age models indicate that the cores used in this study do not contain modern sediments. However, the core-top temperature of CDH 23 is within 0.7°C of the mean annual SST (22.42°C) for the closest World Ocean Atlas [2013] 0.25 degree grid box [WOA 13]. We therefore consider the seasonal bias at our sites negligible [Kienast et al., 2012]. Additionally, a recent synthesis demonstrates that $U^{K'}_{37}$ -estimated SST from the region are within 0.42-1.77°C of mean annual SST [Kienast et al., 2012]. Although many of the alkenone-inferred temperatures are biased towards overestimating observed temperatures in the near-coastal Peru upwelling zone, this bias may be due to large spatial interpolation of limited SST measurements. This is particularly problematic in the complex and spatially heterogeneous near-coastal region examined here.

The timing of SST change across the termination is broadly similar to that suggested by *Koutavas and Sachs* [2008] but the higher resolution of our record provides a new look at the fine-scale structure of the SST evolution across the termination and Holocene. Cool temperatures reconstructed at the last glacial maximum evolved to reach the lowest observed temperature of the entire 25 kyr record, $19.4 \pm 0.1^\circ\text{C}$, about 14.7 ka (14.5-14.9 ka, $p < 0.05$). This temperature decrease is followed by a rapid rise through the late Holocene, peaking to 23.5°C at 6 ka before decreasing again towards present day. This alkenone SST record, and those produced by *Koutavas and Sachs* [2008], differ from EEP Mg/Ca SST records, which instead suggest an early SST warming beginning at least 18 ka [Lea et al., 2000; Pena et al., 2008].

5.2.2 Benthic Temperatures

The reconstructed temperature evolution in the EEP subsurface over the past 25 kyrs varies significantly with depth (Figure 5). Temperatures at 1000 m and 600 m increased by $\sim 3^{\circ}\text{C}$ and 1°C across the termination, respectively. At 1000 m, temperatures were stable at $\sim 2^{\circ}\text{C}$ through the glacial but begin to increase at 17.9 ka (17.8-18.6, $p < 0.05$) by as much as 1°C/kyr . After a brief period of cooling (13-15 ka), warming continued until reaching a maximum temperature (6.7°C) at approximately 11.8 ka. A mid-Holocene maximum occurred at 7.4 ka and is followed by a cooling trend towards present day. The temperature history at 600 m water depth is characterized by centennial-scale temperature swings up to 2.7°C during the late deglacial (12-14 ka). Relative to temperatures at 1000 m, we do not observe significant warming during the deglaciation.

The thermal evolution at the shallowest core site (370 m) along the Peru Margin reveals a similar temperature trend to that seen at 600 m water depth. However, since the 370 m record is truncated at 14.1 ka, we cannot assess the glacial-interglacial temperature change at this site. Instead, we observe a period of highly variable temperature from 12-14 ka, with centennial-scale temperature fluctuations up to $2.6 \pm 0.5^{\circ}\text{C}$, analogous to that seen at 600 m. We also note a near zero offset between temperatures at 600 and 370 m from 14 until 11 ka when a rapid transition at 370 m shifts temperatures to the warmest recorded throughout the record. After, temperatures decreased gradually towards the modern day at an average rate of $0.14^{\circ}\text{C/kyrs}$.

6. Discussion

6.1 Depth dependent response to deglacial forcing in the EEP upper water column

Timing of the initial response to deglacial warming is depth dependent in the EEP upper water column (surface to 1000 m), with subsurface mode and intermediate waters responding thousands of years before the surface ocean. Our deepest site at 1000 m water depth, within the heart of EqPIW, responded first at 17.6 ka (17.6-17.8 ka, $p < 0.05$), synchronously (within the limitations of radiocarbon dating) with rising temperatures over Antarctica. Timing of the initial $\delta^{18}\text{O}$ decrease at 600 m water depth, within SAMW, is more uncertain due to high variability in $\delta^{18}\text{O}$ values across the transition and a less well-constrained age-depth model. However, the best estimate for the timing of the transition is 16.6 ka (16.4-17 ka, $p < 0.05$), approximately 1 kyrs later than the record at 1000 m. Water masses at both sites are sourced primarily from the Southern Ocean where southward shifts in Southern Ocean frontal boundaries, in addition to rising atmospheric temperatures over the Southern Ocean, were likely responsible for the early warming signal in these water masses [e.g. *Denton et al.*, 2010]. At the onset of the deglaciation, SSTs in the circum-Antarctic region warmed in phase with glacier recession in the southern mid-latitudes and southward displacement of the southern westerly winds (SWWs) to at least the northern boundary of the SAMW formation region [*Lamy et al.*, 2007; *Caniupan et al.*, 2011; *Ho et al.*, 2012; *Barker et al.*, 2009; *Putnam et al.*, 2013]. SST records from the SAMW and AAIW water mass source regions rise shortly after 19 ka, in phase with Antarctica and each other [*Lamy et al.*, 2007; *Caniupan et al.*, 2011; *Ho et al.*, 2012; *Barker et al.*, 2009]. We therefore suggest that SAMW and AAIW warmed synchronously and the delay recorded at 600 m is either a statistical fallacy or results

from a change in water mass contributions from southern (SAMW, AAIW), northern (NPIW), and/or subtropical waters (STMW) at the core site, with the subtropical and northern hemisphere sourced waters warming later than those sourced from the southern high latitudes.

Timing of the initial decrease in benthic $\delta^{18}\text{O}$ depends on a combination of factors, including global ice volume as well as the in-situ temperature and salinity properties of seawater. Therefore, the bottom-up timing likely reflects a depth dependence in (1) the timing of the influx of light $\delta^{18}\text{O}$ from the North Atlantic to the Pacific subsurface as suggested by *Stern and Lisiecki* [2014], (2) the response of intermediate and mode water temperature and salinity properties, and/or (3) changes in the fractional water mass contributions to the core site.

The spatiotemporal pattern of deglacial meltwater and propagation of its light $\delta^{18}\text{O}$ signature into the ocean interior remains poorly constrained; model output [Friedrich and Timmermann, 2012] and observations [Stern and Lisiecki, 2014] from the intermediate and deep oceans are in direct opposition, particularly in the Pacific. We therefore assume a globally uniform response in seawater $\delta^{18}\text{O}$ to ice volume changes over the past 25 ka when estimating intermediate depth temperature histories from our benthic foraminiferal $\delta^{18}\text{O}$ records. After applying an ice volume correction [Fairbanks *et al.*, 2005] we still observe the earliest response at depth, with temperatures at 1000 m (EqPIW/AAIW) mirroring those over Antarctica (Figure 5). Sea surface temperature records from SAMW and AAIW formation zones exhibit similar temperature histories to that of EqPIW in the EEP [Lamy *et al.*, 2007; Caniupan *et al.*, 2011; Ho *et al.*, 2012]. The synchronous timing between Antarctic air recorded in EPICA Dome C ice core and

surface ocean temperatures throughout the Southern Ocean suggest SST changed rapidly and in phase with temperatures over the continent (Figure 5) [Jouzel *et al.*, 2007; Lamy *et al.*, 2007; Caniupan *et al.*, 2011; Ho *et al.*, 2012].

Warming at the EEP surface is delayed until 15 ka, approximately 3 kyrs later than subsurface intermediate waters (Figure 5). This vertical pattern of deglacial warming resembles the latitudinal one shown by Shakun *et al.* [2012] where Antarctic temperatures exhibit a 2 kyr lead over the global average. The 3 kyr delay in EEP surface warming, as recorded by alkenones, is thus supported by the upward (or south to north) propagating signature of deglaciation and demonstrates that EEP SSTs were not sensitive to the early rise in southern hemisphere (SH) temperatures or atmospheric CO₂ at the start of the deglacial. This delay could be the result of a zonal adjustment of SST across the Pacific Basin and/or increased upwelling of cool waters from beneath the thermocline in the EEP.

Delayed warming at the surface is consistent with previous alkenone-SST reconstructions from the region [Kienast *et al.*, 2006; Koutavas and Sachs, 2008] but in opposition to Mg/Ca-based reconstructions of SST, which are synchronous with Antarctic air temperatures [Lea *et al.*, 2000; Pena *et al.*, 2008]. The disagreement may be the result of differences in the depth habitat or seasonality of alkenone producers and planktonic foraminifera [Shaari *et al.*, 2014; Timmermann *et al.*, 2014]. Here, we interpret EEP alkenone U^K₃₇ measurements as a recorder of mean annual sea surface temperatures along the northern Peru Margin (see more detailed discussions in Kienast *et al.*, [2012] and Timmermann *et al.*, [2014] in support of this interpretation). Coupled alkenone SST and Mg/Ca records over the last 16 kyrs from site M772-059, just a few

km from our core location, corroborate this interpretation [Nurnberg *et al.*, 2015].

Absolute temperature estimates and the amplitude of SST change reconstructed by shell Mg/Ca and by alkenones are essentially identical during the deglaciation [Nurnberg *et al.*, 2015]. Furthermore, the progression of rising temperatures towards the surface as recorded by our $\delta^{18}\text{O}$ depth series provides independent support for the lag in SST inferred from alkenone paleothermometry (Figure 4).

An important consequence of a delay in surface warming relative to the subsurface is a decrease in the thermal gradient across the upper water column during the glacial termination and thus, assuming constant salinity, a decrease in density stratification between surface and intermediate water masses. However, given the inherent uncertainties in our temperature proxies outlined above, we assess a change in stratification over the top 1000 m of the water column in temperature, $\delta^{18}\text{O}$, and $\delta^{13}\text{C}$ space (Figure 6). Because these records provide similar stratification histories for the EEP (Figure S1), we normalized all records to a late Holocene value and calculated an average offset between the surface and intermediate water records to produce one index for the strength of stratification (Figure 7). Two previously published planktonic stable isotope records measured on *G. sacculifer* from cores V19-28 (2°22'S, 84°39'W) and V21-30 (1°13'S, 89°41'W) are used to represent surface $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values [Koutavas and Lynch-Stieglitz, 2003]. Measurements from all records were linearly interpolated to ages where there are measurements at site V19-28, the lowest resolution record, and normalized to a late Holocene value (2785 yrs ago). $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ gradient anomalies represent an average of the normalized offset between the surface and subsurface records (Figure 6). The thermal gradient anomaly is calculated in the same manner; however, we

use alkenone SST records from CDH 23 and CDH 26 to represent the SST evolution and calculated benthic temperatures from 370, 600, and 1000 m (Figure 6).

The stratification gradient anomaly plotted in Figure 7 represents the average of the temperature, $\delta^{18}\text{O}$, and $\delta^{13}\text{C}$ gradient anomaly calculations. Each of the three gradient anomaly calculations can be viewed separately in Figure 6. Values of the stratification gradient anomaly greater than zero represent times when stratification between surface and intermediate waters was stronger than the late Holocene, while values less than zero are characterized by lower upper water column stratification. Stratification indices based on our temperature (alkenone SST record and foraminiferal $\delta^{18}\text{O}$ derived temperature calculations) and stable oxygen and carbon isotope records (planktonic records from *Koutavas and Lynch-Stieglitz* [2003] and our benthic isotope records) all indicate that stratification was strongest at the end of the last glacial period, decreased rapidly at the onset of Heinrich Stadial 1 (HS1) and the Younger Dryas (YD) and increased towards present day (Figure 6, 7, see also Figure S1 in the supplemental information).

These records support a latitudinal response by water mass to deglacial warming, with Southern Ocean intermediate and mode waters responding first. The physiochemical properties of Southern Ocean intermediate waters (SOIWs) in the EEP were in phase with temperatures in the SE Pacific and over Antarctica; this implies rapid adjustments of the heat content of intermediate waters with atmospheric temperatures surrounding the Antarctic continent and rapid transport into the EEP subsurface. Any deglacial trends in the EEP upper water column driven by changing air-sea interactions were thus overwhelmed by variability in the transport and/or physio-chemical properties of the SOIWs as demonstrated by the delay in surface warming. We therefore suggest that

during the last glacial termination, a period of rapid change in the global radiation balance, remotely forced “bottom-up” rather than local “top-down” forcing drove oceanographic change in the EEP cold tongue.

6.2 EEP intermediate water structure during HS1 and the YD

The last deglaciation was not a smooth transition from glacial to interglacial conditions but was punctuated by three abrupt millennial-scale events: HS1, the Antarctic Cold Reversal (ACR), and the YD. HS1 (14.6-17.6 ka) and the YD (11.7-12.9 ka) are Northern Hemisphere stadial (SH interstadial) events likely caused by NH ice sheet collapse and the subsequent weakening of the Atlantic Meridional Overturning Circulation [McManus *et al.*, 2004]. Despite NH cooling, global temperatures rose steadily throughout these periods in phase with atmospheric CO₂ levels [Lourantau *et al.*, 2010; Shakun *et al.*, 2012]. Several recent studies suggest intermediate water chemistry, formation rate, and volume export to the low latitudes responded to these high latitude events [e.g. Pahnke and Zahn, 2005; Anderson *et al.*, 2009; Pena *et al.*, 2013]. The benthic isotope records presented here from 370 m, 600 m, and 1000 m, sites that straddle the boundary between SAMW and EqPIW/AAIW, provide a rare opportunity to directly assess the response of EEP intermediate water structure to these events. The results indicate that during HS1 and the YD 1) the boundary between these water masses shoaled by at least 230 m during NH stadials and 2) density stratification in the EEP weakened by approximately 10% relative to the last glacial period.

6.2.1 Shoaling of the SAMW/AAIW Boundary

In the EEP, the YD and HS1 are characterized by anomalously small carbon and oxygen isotope gradients across the EEP upper water column. During the YD, benthic

foraminiferal $\delta^{18}\text{O}$ values at 370 m and 600 m are approximately equivalent and are closer to values seen at 1000 m water depth, while carbon isotope values exhibit local minima in each sediment record (Figure 4). We suggest these periods of isotopic convergence across intermediate depths in the EEP reflect a shoaling of the SAMW/AAIW boundary from its modern day position at ~600 m to at least 370 m depth, with numerous centennial-scale shifts. At these times AAIW (or at the very least a more AAIW-like water mass) bathed all core sites, which contributed to a decrease in density stratification between 370 and 1000 m depth.

The modern boundary between SAMW and AAIW lies at approximately 600 m but may shift in response to climatic changes in their source regions. Physical models show that southward displacement of the westerly winds force deepening and thickening of the thermocline, the depth interval containing SAMW and AAIW, in the Southern Hemisphere, [Toggweiler *et al.*, 2006; Russell *et al.*, 2006; Downes *et al.*, 2011; Hain *et al.*, 2014]. During the YD and HS1, reduced SH pole to equator temperature gradients likely forced southward migration of the SWWs, increasing upwelling in the Drake Passage, forcing stronger Ekman upwelling, and increasing formation of Antarctic Surface Water (AASW) that feeds AAIW production [Oke and England 2004; Toggweiler *et al.*, 2006, Russell *et al.*, 2006; Downes *et al.*, 2011]. Radiocarbon, stable isotope, and carbonate ion concentrations from the Southern Ocean support a downward displacement of the AAIW lower boundary during the deglaciation, consistent with model results for a thickening of the AAIW layer [Pahnke and Zahn, 2005; Allen *et al.*, 2015].

Benthic foraminiferal carbon isotope records from our core sites provide evidence for enhanced AAIW contributions to the EEP subsurface during HS1 and the YD, consistent with a thickening of the thermocline layer as observed further south. Today, the $\delta^{13}\text{C}$ content of AAIW in the EEP subsurface is determined by multiple factors, including the carbon chemistry of upwelled Circumpolar Deep Water (CDW), air-sea exchange in the subantarctic zone, and local rates of primary production. During the deglaciation, enhanced upwelling of light $\delta^{13}\text{C}$ in CDW and transport of this signal to the EEP thermocline via SOIW was the primary cause of the deglacial carbon isotope minimum events [*Spero and Lea, 2002; Martínez-Boti et al., 2015*]. Our records corroborate this assertion. Although overprinting by respiratory CO_2 is a concern, alkenone abundances (C_{37} total), a proxy for the rain rate of organic matter, does not show contemporary maxima with carbon isotope minima. In fact, inferred rates of surface primary production reach a broad minimum during the deglaciation at our core site (Figure S2). Furthermore, the timing of carbon isotope minima in the $\delta^{13}\text{C}$ records from the Peru Margin are consistent with that observed in the Pacific sector of the Southern Ocean [*Ninnemann and Charles, 1997*], the influx of $\delta^{13}\text{C}$ depleted SOIW to the EEP thermocline [*Spero and Lea, 2002*], and resurgence of Upper Circumpolar Deep Water (UCDW) upwelling in the Southern Ocean (Figure 4b) [e.g. *Anderson et al., 2009*] and suggest that advection of isotopically light DIC from the Southern Ocean was the primary driver of carbon isotopic change over the past 25 kyrs.

Our data further suggest AAIW shoaled in the EEP water column during HS1 and the YD. Strengthened upwelling, rising atmospheric temperatures, and enhanced freshwater inputs to the AAIW source region may be responsible for the observed

vertical shifts in EEP subsurface water masses during these intervals [*Mashiotto et al.*, 1999; *Moreno et al.*, 2012]. Of these factors, surface salinity values in the Southern Ocean are the least well-constrained. Today, the freshwater balance over the Southern Ocean is regulated primarily by the seasonal position of the SWWs, which are the main source of freshwater to the Southern Ocean [*Toggweiler et al.*, 2006; *Moreno et al.*, 2012]. During the last glacial termination, poleward shifts in the SWWs likely altered the latitudinal distribution of precipitation, with zones of maximum precipitation tracking poleward movement of the SWWs during the YD and HS1 [*Lamy et al.*, 2010; *Moreno et al.*, 2012].

Sea ice also plays an important role in the fresh water balance over the Southern Ocean. During the deglaciation, sea ice retreat began in phase with SH warm events and was amplified by poleward shifts in the position of the SWWs and the subsequent break-up of sea ice via northeastward Ekman transport [*Gersonde et al.*, 2005; *Levermann et al.*, 2007]. Without sea ice capping its surface, AASWs began to directly exchange heat and fresh water with the atmosphere, which led to greater interaction between the regional hydrologic cycle and the surface ocean and thus enhanced fresh water inputs to AASWs and eventually SOIW [Keeling and Stephens, 2001]. Consequently, during HS1 and the YD, net precipitation to the surface ocean in addition to melting sea ice led to a net increase in the freshwater inputs to the surface ocean and positive buoyancy forcing for AAIW and SAMW [Keeling and Stephens, 2001]. Increased freshwater inputs likely contributed to greater AAIW buoyancy at these times forcing AAIW, and therefore EqPIW, to shoal within the water column. Thus some of the observed isotopic depletion observed in our records during the YD and HS1 likely records freshening as well as

warming of intermediate water masses. Records of seawater $\delta^{18}\text{O}$ from the Pacific sector of the Southern Ocean, however, indicate minimal influence from local hydrologic conditions [Mashiotta *et al.*, 1999]. Regardless, even if freshening did have a significant effect on $\delta^{18}\text{O}$ values at our site, the interpretation is the same: sub-surface density stratification was weak throughout the deglaciation, with pronounced minima during the YD and HS1.

6.2.2 EEP de-stratification events and the global carbon cycle

Stratification across the upper water column was reduced during the last deglaciation relative to the end of the last glacial and late Holocene but reached minimum observed values during HS1 and the YD (Figure 7). At these times, the $\delta^{18}\text{O}$, $\delta^{13}\text{C}$, and temperature gradients across the upper water column decreased by approximately 10% relative to glacial conditions. Although a 10% reduction is not enough for AAIW or even SAMW to mix completely with surface waters, such a reduction in the upper ocean density gradient must have allowed for more vertical mixing and therefore greater penetration of subsurface mode and intermediate waters to the surface mixed layer. This notion is corroborated by a recent study: *Nurnberg et al.*, [2015] demonstrate that temperature gradients were reduced between the sea surface and ~200 m ($\pm$ ~100m), a key depth interval not sampled by our cores, during the deglaciation relative to the Holocene at a site within a few kilometers of sites CDH 23 and CDH 26, presented here. In addition, three proxy records of surface ocean conditions—the alkenone $\text{C}_{37}\text{total}$, planktonic foraminiferal $\delta^{13}\text{C}$ values, and alkenone SSTs—support an increase in communication between intermediate and surface waters during these millennial events.

Productivity in the EEP depends primarily on the supply of cold, nutrient-rich intermediate waters from beneath the thermocline and iron limitations [*Sarmiento et al.*, 2004; *Pennington et al.*, 2006]. Therefore, greater productivity suggests enhanced nutrient delivery from the subsurface to the surface, either via stronger vertical mixing or higher nutrient concentrations in the upwelled water. During the deglaciation, productivity, as inferred from our alkenone C_{37} total record of coccolithophorid productivity, was lower than during the glacial and Holocene at our site, but we find evidence for small increases in primary productivity during both HS1 and the YD (Figure 6). We observe local maxima in coccolithophore (C_{37} alkenone) [Figure 7] and diatom (brassicasterol) [Calvo et al., 2011; Pena et al., 2013] productivity in addition to planktonic isotope minima [*Spero and Lea*, 2002]; all three are independent indicators of enhanced primary production and greater nutrient delivery to the surface ocean. Reductions in the vertical density gradient over the upper water column observed during these intervals suggest strengthened vertical mixing was responsible for the small increases in surface primary production. During these intervals SOIWs composed as much as 25% of EUC water, roughly a 20% increase from the Holocene and present day average [Pena et al., 2013]. Finally, greater penetration of cool intermediate waters to the sea surface could help explain how SSTs remained cool despite rising atmospheric CO_2 concentrations.

Reduced upper ocean stratification during the YD and HS1 was not limited to the EEP but is also observed in the southwest and southeast Pacific (Figure 6) [*Bostock et al.*, 2004; *Siani et al.*, 2013]. The analogous timing and nature of the transition from strong glacial stratification to weak deglacial stratification in all records is consistent with a

change in upper water column stratification forced by the same mechanism at each site: an early response of subsurface intermediate waters to SH climate at the end of the last ice age [*Bostock et al.*, 2004; *Pena et al.*, 2013; *Siani et al.*, 2013]. Together these records suggest a South Pacific-wide decrease in water column density stratification during HS1 and the YD.

Given the vastness of the South Pacific, these de-stratification events may be responsible for ventilating large amounts of CO₂ from a deep Southern Ocean reservoir [*Ninnemann and Charles*, 2002; *Lourantou et al.*, 2010]. Ice core records from Antarctica show a two-step increase in atmospheric CO₂ during the YD and HS1, in phase with periods of decreased stratification in the EEP and the South Pacific (Figure 7) [*Lourantou et al.*, 2010, *Marcott et al.*, 2014]. These findings are consistent with a rapid decrease in stratification in the Southern Ocean, enhanced upwelling of carbon-rich UCDW and degassing of isotopically light remineralized organic carbon to the atmosphere. A south Pacific-wide “carbon window”, however, requires that upwelled UCDW was resubducted as SOIW faster than the air-sea carbon equilibration time or became entrained into SAMW and/or AAIW before it reached the surface [*Holte et al.*, 2012; *Holte et al.*, 2013]. This notion is corroborated by marine radiocarbon evidence from the southwestern Pacific Ocean, which showed that—at least at the onset of HS1—upwelled carbon-rich UCDW did not fully equilibrate with the atmosphere for at least 1000 years [*Rose et al.*, 2010]. Instead, the non-equilibrated CO₂ was likely resubducted or entrained in AAIW and released to the atmosphere somewhere along the flow path between the Southern Ocean and the EEP; decreased density gradients over the S. Pacific upper water column

allowed the resequenced CO₂ to mix upwards, resurface and exchange with the atmosphere.

Boron isotopes provide direct evidence for CO₂ outgassing far afield from the Southern Ocean in the sub-Antarctic Atlantic and at least 3 sites across the equatorial Pacific, including the EEP [Palmer and Pearson, 2003; Douville *et al.*, 2010; Kubota *et al.*, 2014; Martínez-Boti *et al.*, 2015]. At site ODP 1238, ~300 km northwest of our site, Martínez-Boti *et al.* [2015] record CO₂ outgassing synchronous with periods of reduced upper ocean stratification. Although correlation does not imply causation, the remarkable similarity between records suggests reduced stratification enhanced delivery of CO₂-rich subsurface waters to the surface mixed layer, providing a pathway for oceanic carbon to the atmosphere (Figure 7). Finally radiocarbon and $\delta^{13}\text{C}$ depleted benthic and planktonic foraminifera from the Galapagos platform and the Peru Margin during HS1 and the YD provide further support for an inflow of carbon-rich intermediate waters to the EEP [Spero and Lea, 2002; Stott *et al.*, 2009; Chapter 3].

7. Summary and Conclusions

Benthic foraminiferal isotope and temperature records recovered from three intermediate water depths in the EEP provide evidence for large changes in upper water column structure during the deglaciation forced by climate change over the Southern Ocean [Figure 8]. Rising temperatures over Antarctica led to poleward shifts in the westerly wind belt that together altered the buoyancy and water mass boundaries of southern ocean intermediate waters. SOIWs responded first, synchronous with atmospheric temperatures over the Antarctic continent, and transported the early SH deglacial warming response along its flow path to the base of the equatorial thermocline

and the EEP. Surface warming in the EEP was delayed by 3 kyrs, while records from the shallow subsurface at 600 m, document an averaged time response of the two, ~1 kyrs after Antarctica.

The observed upward-propagation in our benthic $\delta^{18}\text{O}$ and temperature records does *not* favor the tropical surface ocean as an early responder and promoter of deglaciation. Instead, the EEP surface ocean was likely a passive responder to subsurface change driven from the bottom-up by the southern high latitudes. During SH warm events (YD/HS1), the early warming response of AAIW drove rapid reductions in vertical temperature and salinity gradients in the EEP upper water column, which led to reduced stratification, shoaling of water mass boundaries, and enhanced vertical mixing. Thus despite rising atmospheric CO_2 concentrations, greater penetration of cool subsurface waters to the sea surface during these events helped maintain a strong cold tongue.

Reduced upper ocean stratification during the YD and HS1 was not limited to the EEP but is also observed in records from the southwest and the southeast Pacific [*Bostock et al.*, 2004; *Siani et al.*, 2012]. Comparable timing and structure of these records suggest an early rise in SOIW heat content, synchronous with SH temperatures, drove change in upper water column density gradients across the entire south Pacific [*Bostock et al.*, 2004; *Siani et al.*, 2013]. These gyre-wide de-stratification events are in phase with rising atmospheric CO_2 concentrations and increasingly depleted atmospheric carbon isotope signatures [*Monnin et al.*, 2001; *Lourantau et al.*, 2010]. We suggest a decrease in stratification across the upper water column led to enhanced vertical mixing, which provided a pathway for carbon-rich intermediate waters to invade surface waters and exchange with the atmosphere during the YD and HS1.

The response of intermediate and mode waters to rapid global warming during the last deglaciation may provide a good basis for predicting their response to modern climate change. Both periods of rapid global warming are associated with shifts in the position and strength of the southern westerly winds. Today movement of the SWWs is associated with the polarity state of the Southern Annular Mode, which has shifted in response to ozone depletion and warming global temperatures over the last few decades. These shifts had significant effects on intermediate water formation rates and properties, perhaps comparable to those observed in the past [*Naveira-Garabato et al.*, 2009; *Thompson et al.*, 2011, *Schmidtko and Johnson*, 2012]; since 1925 AAIW core densities decreased resulting in an overall shoaling of the water mass by 30-50 dbar decade⁻¹, with the larger shifts occurring towards the southernmost extent of AAIW [*Schmidtko and Johnson*, 2012]. These rates of warming within AAIW are unprecedented in the historical record [*Schmidtko and Johnson*, 2012]. Based on our observations from the deglaciation, we therefore predict reduced stratification and enhanced vertical mixing to occur, with the potential to alter the accumulation rate of carbon in SOIWs, key sinks for anthropogenic carbon today [*Sabine et al.*, 2004; *Schmidtko and Johnson*, 2012].

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Figures:

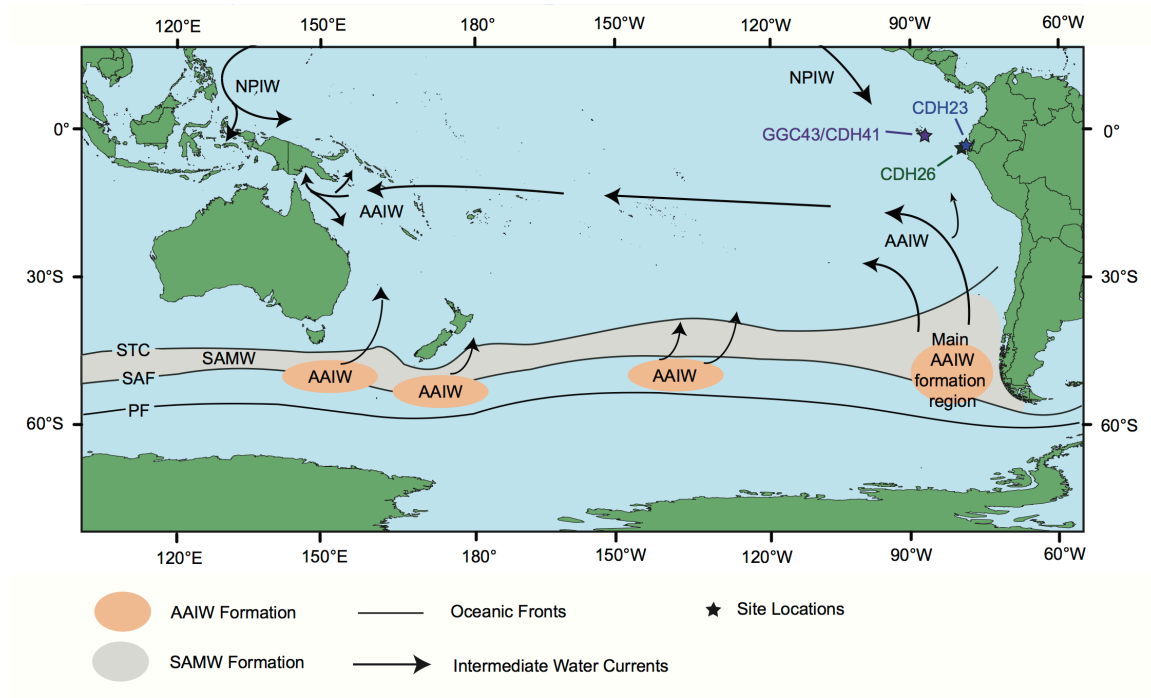


Figure 1. (a.) Key formation regions of AAIW and SAMW and their subsurface paths to the EEP. In the Pacific, AAIW forms primarily in the SE Pacific along the tip of South America, just northwest of the Drake Passage, with small contributions from west and east of New Zealand and the East Pacific Rise [Bostock et al., 2013]. SAMW forms along the Subantarctic Front (SAF) south of the Subtropical Convergence (STC). AAIW, SAMW, and small contributions from NPIW contribute to Equatorial thermostad waters.

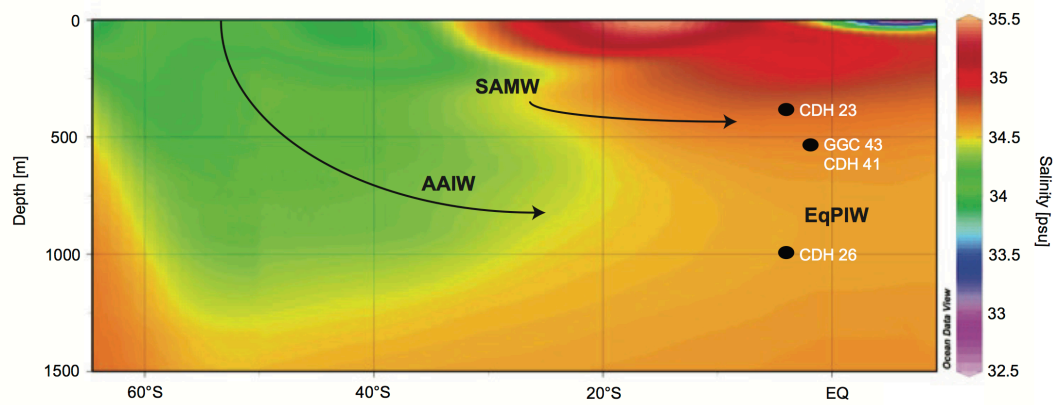


Figure 2. Salinity of the EEP upper water column from 65°S to 10° N averaged along 87.5 °W [WOA 13]. AAIW is characterized by a core of low salinity water at intermediate depth, with psu values between 34.3 and 34.5. SAMW is saltier but warmer than AAIW. EqPIW is composed of a mixture of AAIW and Pacific Deep Water [Bostock et al., 2010, 2013]. Cores used in this study are plotted at their depth and latitude of recovery. The deepest core, CDH 26, at 1000 m lies within EqPIW, GGC43 and CDH41 lie along the boundary between the water masses, and CDH23 is bathed by SAMW.

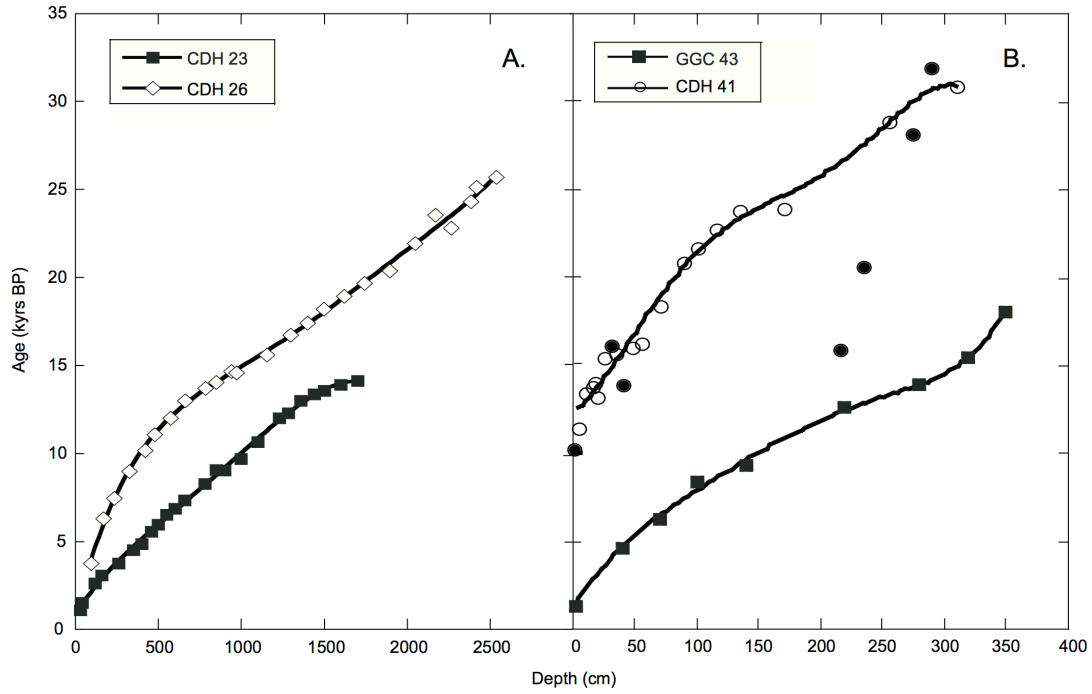


Figure 3. Age models for the four cores used in this study. The age model for GGC 43 is based on 9 radiocarbon measurements of the planktonic foraminifera, *Globigerinoides ruber*, (filled squares) and 17 of 24 measurements from CDH41 (open circles). CDH 23 and CDH 26 are better constrained with 24 and 25 AMS ^{14}C measurements, respectively, made on the thermocline dwelling foraminifer, *Neogloboquadrina dutertrei*. AMS ^{14}C ages were converted to calendar ages using the calibration curve of Fairbanks et al. [2005]. A constant reservoir age of 500 years is assumed for all measurements. Seven AMS ^{14}C measurements from CDH 41 were removed from age model calculations (filled circles).

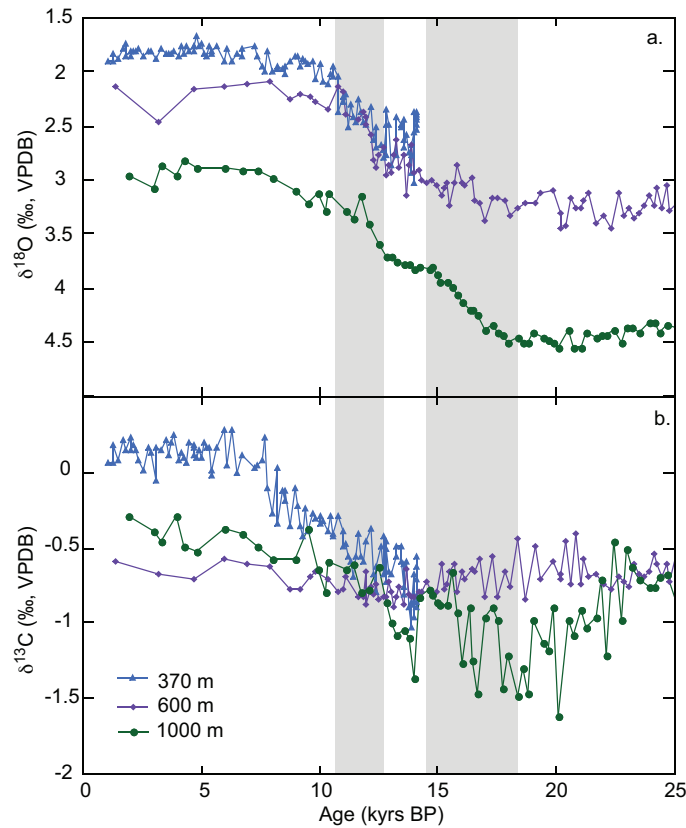


Figure 4. Stable isotope records from the benthic foraminifer, *Uvigerina peregrina*, from three water depths, 370 m, 600 m, and 1000 m, in the EEP. (a.) Oxygen isotope values converge between 12.3 and 14.3 ka at 370 and 600 m. (b.) Carbon isotope values at 370, 600, and 1000 m also converge at this time. Carbon isotope minima occur between 12.7-14.2 and 15.6-22 ka. Because *Uvigerina peregrina* is an infaunal species the recorded values may in part reflect porewater chemistry and thus the accumulation rate of organic carbon [Zahn et al., 1986]. We therefore interpret these data with caution, noting that the excursions to large negative values likely reflect an influx of ^{13}C depleted carbon to the EEP subsurface in intermediate water as well as pore water production of depleted ^{13}C via high productivity and organic matter export to the sediments.

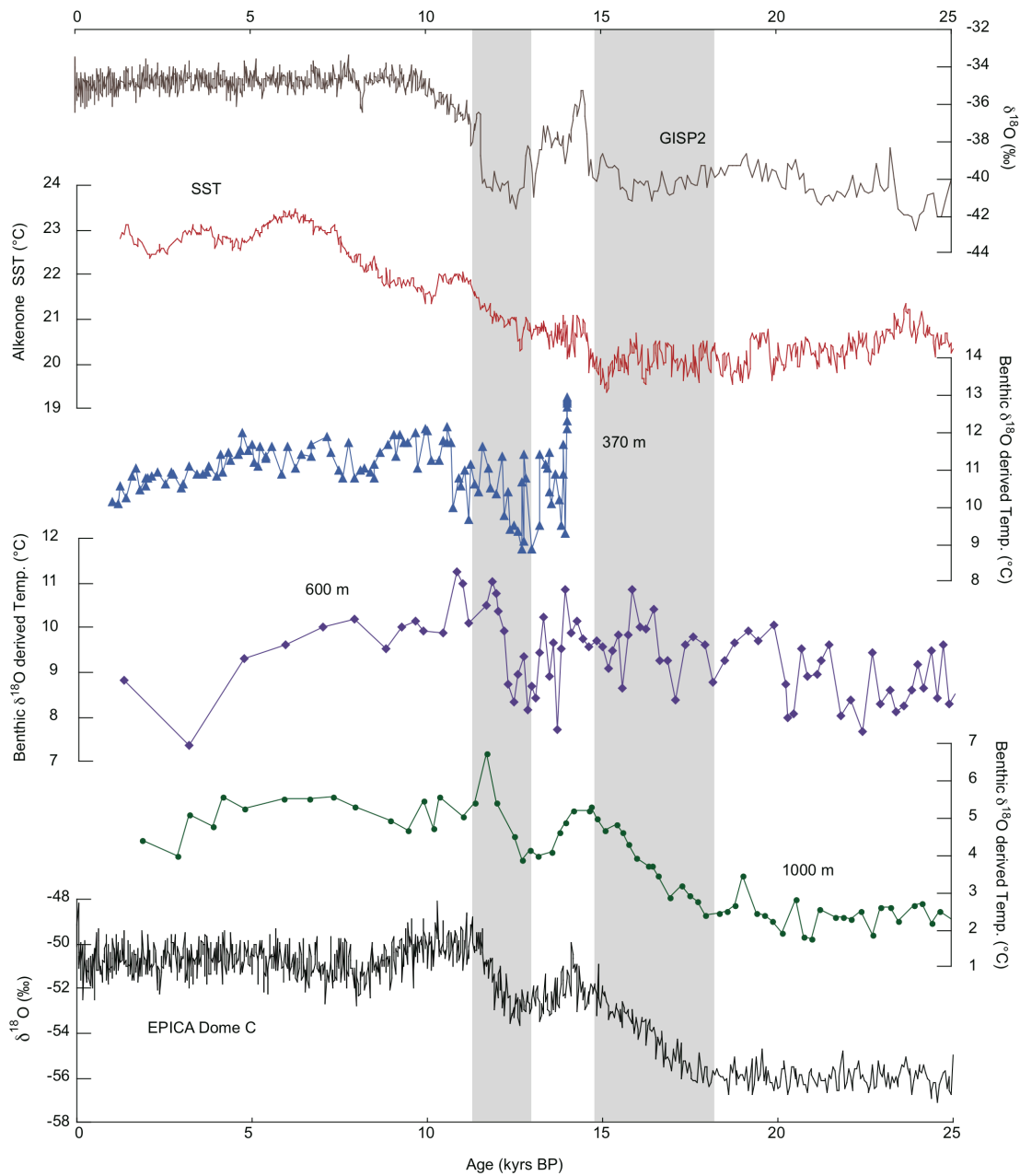


Figure 5. Comparison between temperature records from the EEP and high latitude records of global change during the last deglaciation. Ice core records from Greenland (top, GISP2) and Antarctica (bottom, EPICA Dome C) document the temperature evolution in the Northern and Southern Hemisphere, respectively [Grootes et al. 1993;

Jouzel et al. 2007]. The subsurface temperature record from 1000 m appears to have SH timing for deglacial warming. At 600 m, we do not observe a deglacial warming trend, which is likely explained by a mean shoaling of the SAMW/AAIW boundary to at least 600 m during SH warm events. At the sea surface, warming is delayed until 14.7 ka.

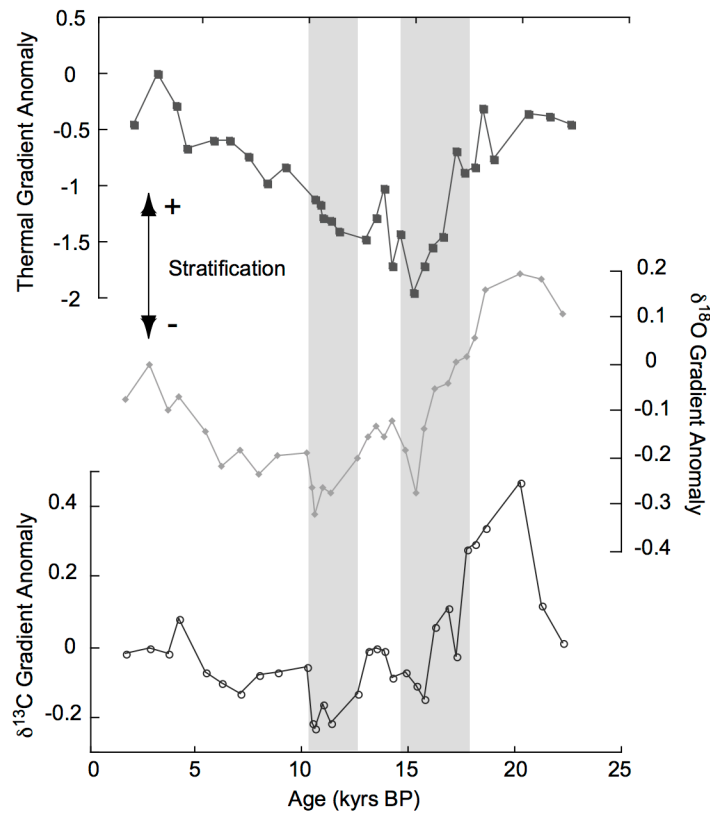


Figure 6. The stratification gradient anomaly was calculated as the offset between records from the sea surface, 370 m, 600 m, and 1000 m in $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ and temperature space. Subsurface temperature records were linearly interpolated to ages where there are $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ measurements in the lower resolution surface record from site V19-28. All records were subsequently normalized to a late Holocene value (2785 yrs ago). $\delta^{18}\text{O}$, $\delta^{13}\text{C}$, and thermal gradient anomalies represent an average of the normalized offset between the surface and subsurface records. The stratification gradient anomaly represents the average of these three gradient anomaly calculations.

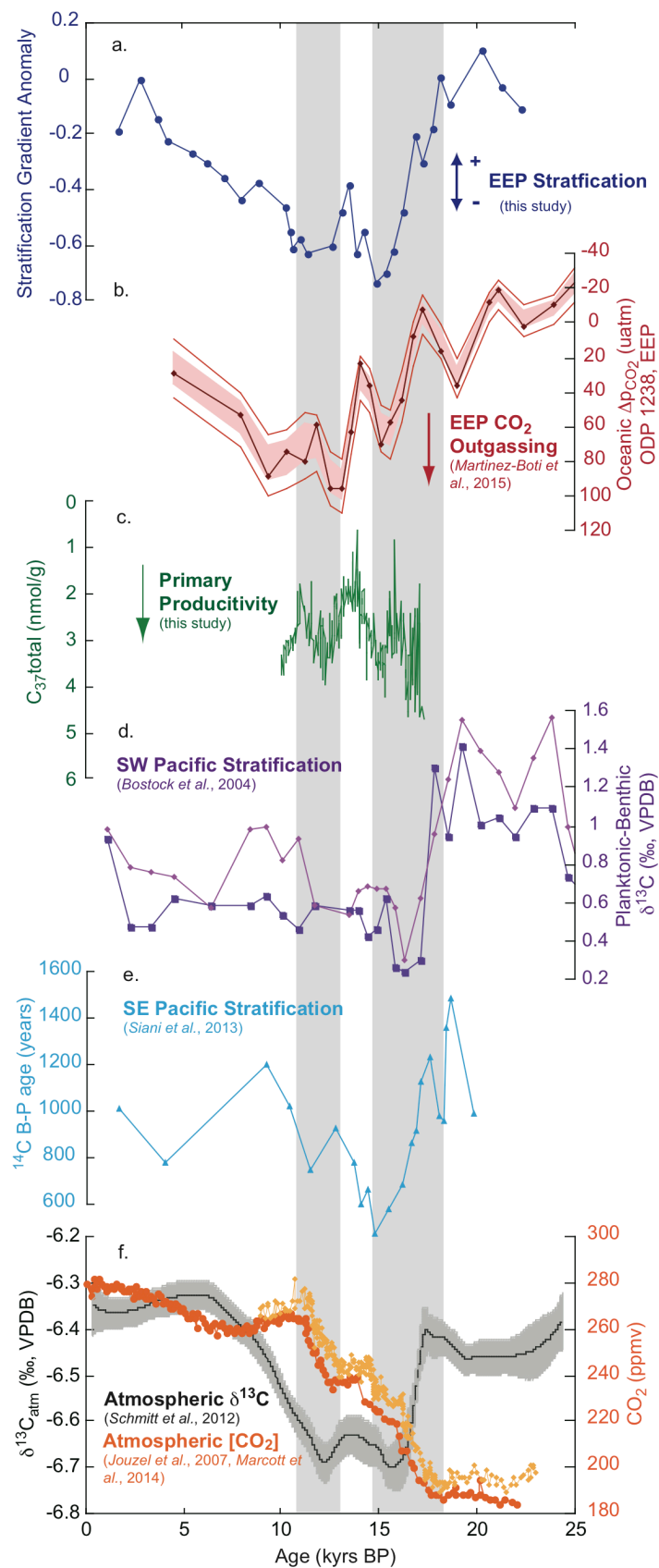


Figure 7. Measures of stratification in (a) the EEP, (d), the northwest Tasman Sea, southwest Pacific [Bostock *et al.*, 2004], and (e.) the southeast Pacific [Siani *et al.*, 2013]. (b.) Surface ocean $\Delta p\text{CO}_2$ at ODP site 1238 in the EEP [Martinez-Boti *et al.*, 2015]. (c.) Coccolithophore productivity inferred from alkenone abundances ($\text{C}_{37}\text{total}$) at sites CDH 23 and CDH 26 along the northern Peru Margin [this study]. (e.) Atmospheric $\delta^{13}\text{C}$ record from Antarctic ice cores [Schmitt *et al.*, 2012] (f.) Atmospheric CO_2 concentrations from Epica Dome C [Lourantau *et al.*, 2010] and the WAIS Divide core [Marcott *et al.*, 2014]. Stratification in the EEP decreased at the onset of the last glacial termination, 18 ka, synchronous with oceanic CO_2 outgassing to the atmosphere in the EEP [Martinez-Boti *et al.*, 2015], increasing coccolithophore productivity along the Peru Margin [this study] and rising atmospheric CO_2 levels [Lourantau *et al.*, 2010; Marcott *et al.*, 2014]. Periods of reduced stratification at all Pacific sites are synchronous with each other, which suggests that reduced stratification created a pathway for isotopically light carbon to escape from the deep ocean to the atmosphere during HS1 and the YD.

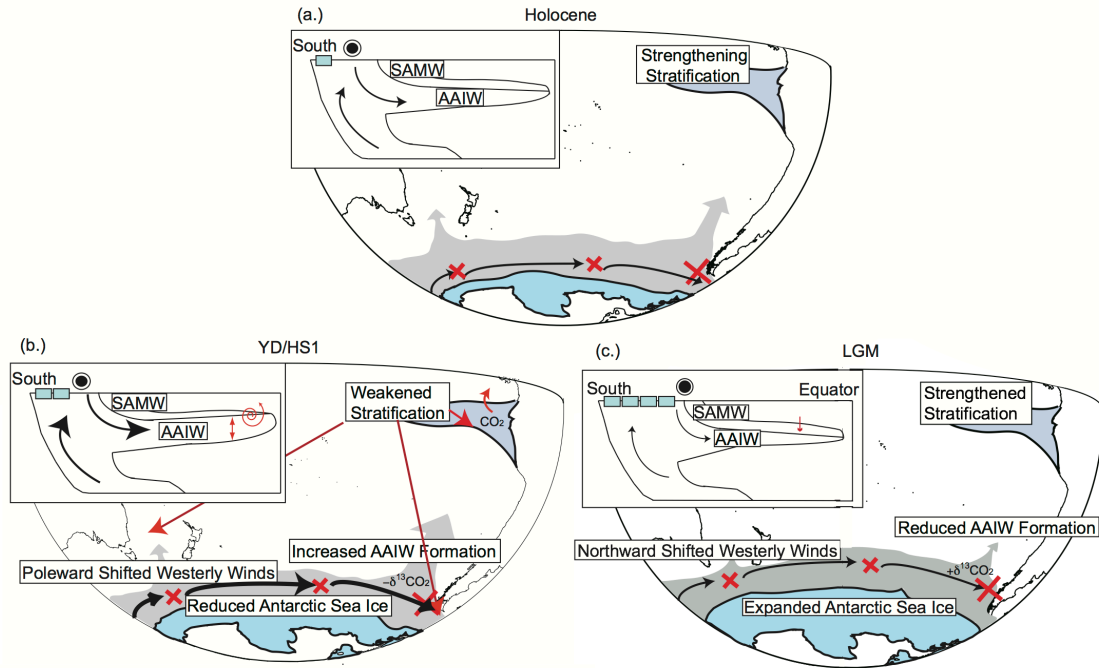


Figure 8. Connections between the southern high latitudes and the EEP during (a.) the Holocene, (b.) the YD/HS1, and (c.) the LGM. AAIW formation regions are shown with red x's. For each time slice, a 2-d north-south transect shows structural change in intermediate and mode water geometry. AAIW is thinner and deeper in the water column during the LGM and shoals and thickens during the YD/HS1 in response to changing sea ice extent, shifts in the position and strength of the SWWs, and rising atmospheric temperatures.

Supplementary Information

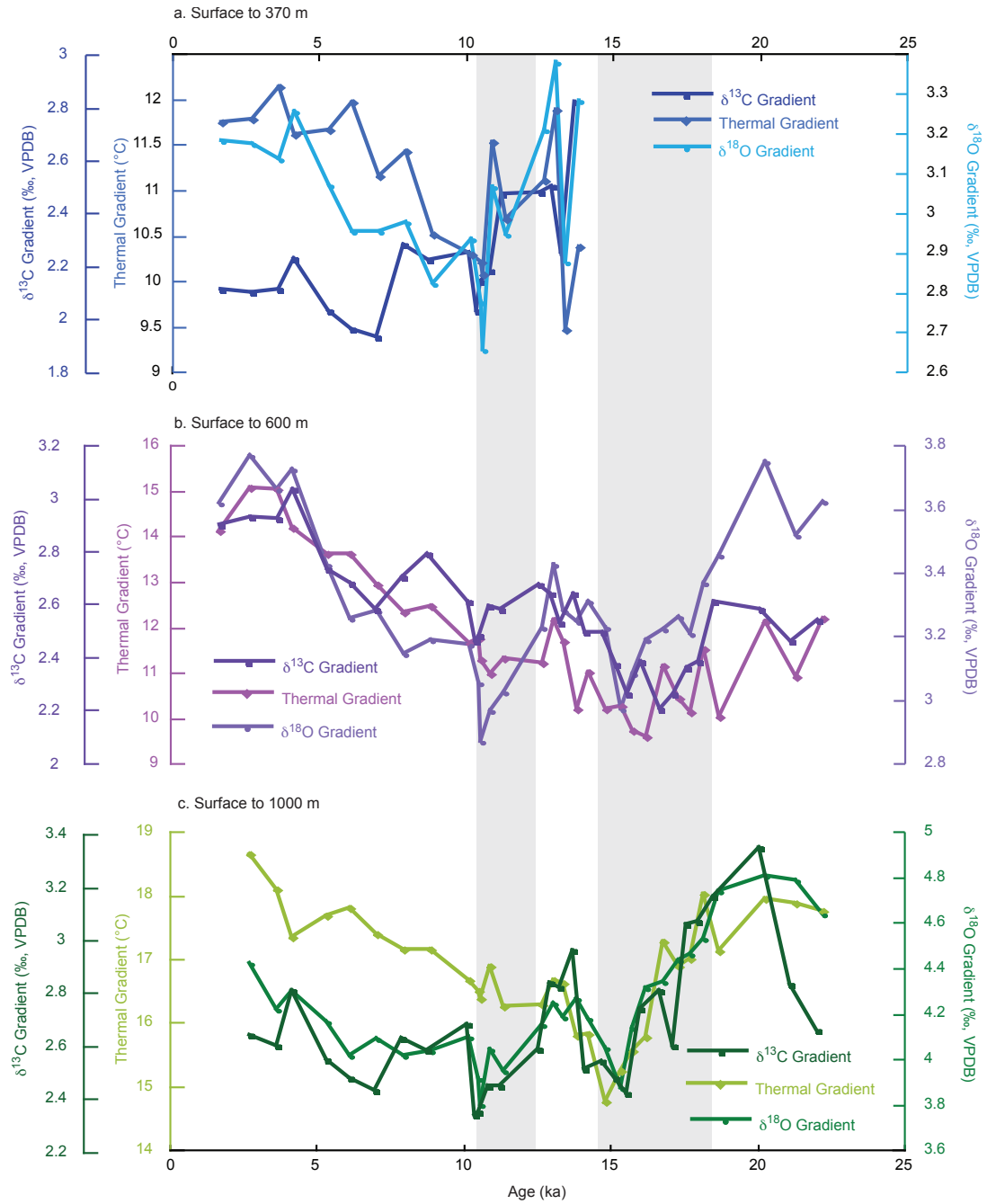


Figure S1. The stratification gradient in $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ and temperature space was calculated as the offset between records from the sea surface, 370 m, 600 m, and 1000 m. Here we show the $\delta^{18}\text{O}$, $\delta^{13}\text{C}$, and temperature gradient between the surface records and each

subsurface record individually. (a.) $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ and thermal gradient between the surface and 370 m, (b.) $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ and thermal gradient between the surface and 600 m, (c.) $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ and thermal gradient between the surface and 1000 m. Calculated gradients to 370, 600, and 1000 m each exhibit reduced stratification during the deglaciation, with the smallest gradients occurring during the YD and HS1 (gray bars).

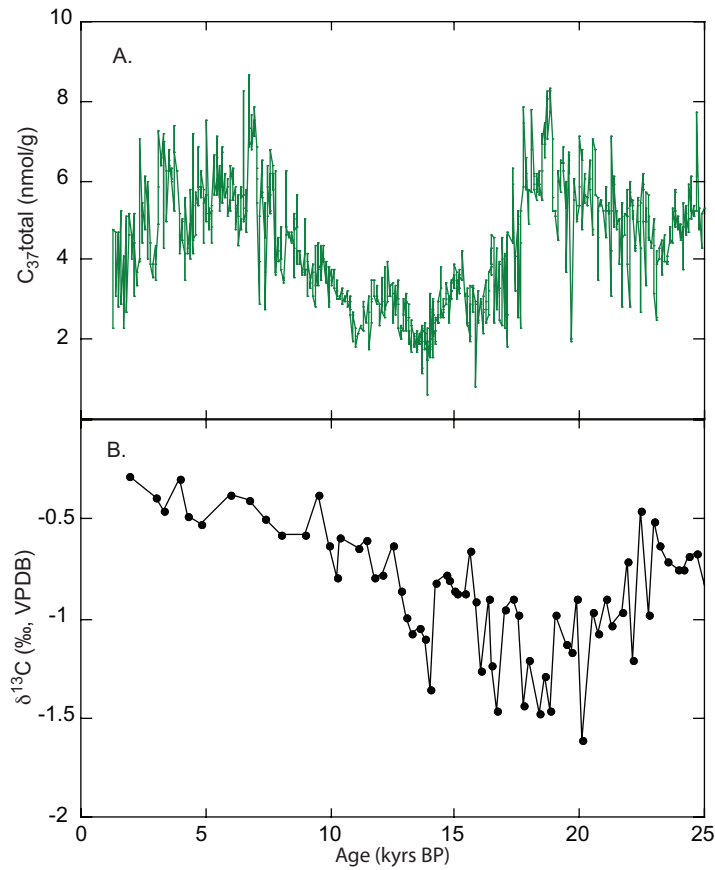


Figure S2. (A.) Alkenone abundances ($C_{37}total$) at sites CDH 23 and CDH 26 along the northern Peru Margin. (B.) Carbon isotopes measured on the benthic foraminifer *Uvigerina peregrina* at 1000 m water depth at site CDH 26. The $C_{37}total$, a proxy for the rain rate of organic matter, is not correlated to subsurface $\delta^{13}C$ values, which suggests overprinting by respiratory CO_2 is not the primary control on *Uvigerina* $\delta^{13}C$ values. Instead, advection of isotopically light DIC from the Southern Ocean was likely the primary driver of carbon isotopic change at this site during the last deglaciation.

CHAPTER TWO

Rapid variations in deep ocean temperature not unprecedented in the Holocene

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Abstract

The observational record of deep-ocean variability is short, which makes it difficult to attribute the recent rise in deep-ocean temperatures to anthropogenic forcing. Here, we test a new proxy – the oxygen isotopic signature of individual benthic foraminifera – to detect rapid (i.e. monthly to decadal) variations in deep-ocean temperature and salinity in the sedimentary record. We apply this technique at 1000 m water depth in the Eastern Equatorial Pacific during seven 200-year Holocene intervals. Variability in foraminifer $\delta^{18}\text{O}$ over the past 200 years is below the detection limit, but $\delta^{18}\text{O}$ signatures from two mid-Holocene intervals indicate temperature swings $>2^\circ\text{C}$ within 200 years. Transport between the surface and deep ocean operating on human timescales or natural unforced variability, not active during the historical record, are potential explanations. Distinguishing externally forced climate trends in deep ocean properties from unforced variability should be possible with systematic analysis of suitable deep-sea cores.

1. Introduction

On centennial and shorter timescales the deep ocean (>1000 m water depth) is often considered a passive observer of global change. Velocities in the deep ocean are slow, orders of magnitude lower than in the surface ocean, and eddies are also thought to be weaker at depth [Lozier, 2010]. As a result, most regions below 1000 m have not been ventilated, i.e., have not visited the sea surface to exchange heat, carbon, or other conserved properties with the atmosphere, for hundreds of years [Gebbie and Huybers, 2012]. Variability in stratification can be communicated more rapidly by waves or occur via hydrodynamic instabilities, neither of which require direct ventilation [Masuda *et al.*,

2010; *LaCasce and Pedlosky*, 2004]. Recent hydrographic sections taken under the GO-SHIP program provide evidence for quick transport of surface anomalies to the deep ocean, with rising temperatures below 700 m on the order of 0.1°C/decade and accelerated warming over the past 15 years consistent with the recent slowdown in rising mean surface air temperatures [*Purkey and Johnson*, 2010]. Because the temperature rise in the deep ocean parallels the observed surface warming in many places, this suggests that it is capable of exchanging heat and freshwater on decadal to centennial timescales.

Rates of warming in the deep ocean since 1999 are unprecedented in the historical record [*Balmaseda et al.*, 2013], but the record is short (<135 yrs) [*Hobbs and Willis*, 2013] and sparse, and therefore provides potentially insufficient constraints on the natural variability of the deep ocean. Repeat hydrographic sections below 700 m are rare [*Roemmich et al.*, 2012; *Arbic et al.*, 2014], which limits the degree to which variability and trends can be associated with particular processes. As a result, spatial variability is often used to estimate temporal variability, but many processes that contribute to deep ocean variability prevent simple conversion of observed spatial variability to temporal variability [*Arbic et al.*, 2014]. Sampling has improved since 2004 with the Argo float program, but the record is short and limited in depth.

Here, we test a new tool to examine the natural variability of intermediate and deep ocean temperature change on short time scales beyond the historical record. Specifically, we measured the stable oxygen isotopic signature of individual benthic foraminifera from ocean sediments to reconstruct monthly to decadal-scale variability in deep ocean temperature and salinity. We apply this technique to two sediment cores, CDH-26 (03°59.16 S, 81°18.52 W) and MC-18A (03°59.20 S, 81°18.60 W), collected in

2009 from the eastern equatorial Pacific (EEP) at 1000 m water depth, approximately the boundary between deep and intermediate water masses in the eastern tropical Pacific (Fig. 1, supplemental information, Figure S1). Benthic foraminiferal shell chemistry therefore reflects the mixed influence of intermediate and deep ocean water masses; Equatorial Pacific Intermediate Water (EqPIW), a mixture of Antarctic Intermediate Water (AAIW) and Pacific Deep Water, bathes the core sites today. Flow is generally westward between 500 and 1500 m at 2 to 4°S and eastward at greater depths [Firing *et al.*, 1998]. AAIW is the dominant component of EqPIW and is the main conveyor of heat and salt to the core site, ageing 750 years before reaching the EEP subsurface [McCartney, 1977; Sloyan and Rintoul, 2001; Bostock *et al.*, 2010; 2013; Gebbie and Huybers, 2012; Trenberth and Fasullo, 2014]. Waters at the site therefore have not been at the sea surface to exchange heat and gas with the atmosphere since well before industrialization. Figure 1 indicates the level of variability near the study site as a function of depth and location. Interestingly, in this location the whole span of observed modern variability is too weak to detect, but two episodes of variability within the Holocene are sufficiently strong to detect.

2. Material and Methods

The average oxygen isotopic signature measured on multiple, typically 2-4, benthic foraminiferal shells is an established proxy for the mean temperature and salinity of deep water masses through time [e.g. Shackleton, 1974]; the $\delta^{18}\text{O}$ signature of shell carbonate is determined by (1) the $\delta^{18}\text{O}$ of ambient seawater (ice volume and salinity) and (2) seawater temperature during shell growth. By measuring individuals rather than bulk samples, we exploit the short lifetime of benthic foraminifera, ~ 1 month, to

reconstruct the variability rather than the mean of deep ocean physical properties [Murray, 1991; Koutavas et al., 2006]. The benthic foraminifer, *Uvigerina Peregrina*, is abundant in both cores and is therefore used exclusively for this study. *U. peregrina* live in the sediment, inhabiting the top 2 cm of the sediment. Studies show *U. peregrina* calcifies in approximate equilibrium with the overlying seawater and provides reliable records of seawater $\delta^{18}\text{O}$ and temperature through time [e.g. Shackleton, 1974; Fontaneir et al., 2006]. Stable isotopic signatures of live infaunal benthic foraminifers, including *U. peregrina*, from the Bay of Biscay indicate that microhabitat does not have a systematic affect on test $\delta^{18}\text{O}$ but does affect test $\delta^{13}\text{C}$ signatures [Fontaneir et al., 2006]. We therefore examine the relationship between test $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$; we find that foraminifers with anomalous $\delta^{13}\text{C}$ values are not more likely to have anomalous $\delta^{18}\text{O}$ values (Supplemental Information, Text S2).

Seven approximately 200-yr intervals were chosen to assess deep water variability. These intervals span -56-153 yrs BP, 3190-3343 yrs BP, 3406-3580 yrs BP, 4008-4151 yrs BP, 6104-6284 yrs BP, 6910-7095 yrs BP and 7912-8095 yrs BP but will be named by midpoint age (Table 1). The corresponding core depth ranges span up to 20 cm in the multicore to just 6 cm in the late Holocene portion of CDH-26 based on calculated sedimentation rates from AMS ^{14}C constrained chronologies (Supplemental Information, Figure S2) [Bova et al., 2015]. CDH-26 does not contain modern sediment. We therefore estimate variability over the past 200 years using sediment from MC-18A. Both cores exhibit high sedimentation rates, ~ 100 cm/kyrs, likely due to the nearby outflow of the Guayas River. Each interval was sampled at 1 cm resolution and wet-sieved to separate size fractions. Forty individual *U. Peregrina* were picked from the

300-355 μm size fraction uniformly across each interval and measured on a Finnigan MAT 252 isotope ratio mass spectrometer with a Kiel III carbonate device. Prior to analysis, all individuals were sonicated for 10 seconds in 50 μL of ethanol to remove unwanted material from shells and weighed individually. All foraminifera were photographed at 60x magnification except 20 individuals from the intervals spanning 3190-3343 yrs BP and 4008-4251 yrs BP (Supplemental Information, Figure S5). In addition to the sampled foraminifera, 102 standards of known oxygen and carbon isotopic values ranging in weights from 4 μg to 111 μg were run to assess machine linearity and sample size dependent precision (Supplemental Information, Figure S3).

4. Results

The small size of individual foraminifers (16 μg - 87 μg) is problematic and reduces analytical precision relative to conventional analyses. As a consequence, the average analytical error for small samples is nearly 0.15‰, over double that for standard sized samples and gets larger (smaller) with decreasing (increasing) sample size. This is apparent in the small standard dataset (supplemental information, Figure S3). In total we discard 21 standards and 52 samples. For a discussion of the controls on data exclusion please see the Supplemental Information (Text S1).

Once data quality controls are enforced we calculate the variance and range in each time interval and perform a Kolmogorov-Smirnov Test to determine whether $\delta^{18}\text{O}$ distributions about the mean within each sampling interval are significantly more variable than a distribution of lab standards. Failure to reject the null hypothesis indicates that variability is explained by analytical noise alone. Each sample distribution was compared to 5000 randomly sampled distributions of standard measurements (with replacement).

Sample size was taken into account by only comparing sets of standards that are indistinguishable based on sample size (Supplemental Information, Text S1). Results from the size-distribution matched K-S tests indicate that only the sampling intervals at 4080 ($p < 0.01$) and 7003 yrs BP ($p < 0.10$) are distinguishable from the standard with confidence (Fig. 3). They are also distinguishable from the $\delta^{18}\text{O}$ distribution from the past 200 years ($p < 0.01$). The statistics for each interval and a lab standard appear in Table 2.

5. Causes of Foraminiferal $\delta^{18}\text{O}$ Variability

High variability intervals are distinguished by a handful of anomalous $\delta^{18}\text{O}$ values that create tails in their distributions towards either positive or negative values (Figure 2, 3). These anomalous $\delta^{18}\text{O}$ signatures can be explained in one of three ways. First, individual benthic foraminifera may be inconsistent recorders of deep water temperature and salinity. If true, this challenges the reliability of countless benthic $\delta^{18}\text{O}$ records used to infer paleo-temperatures and past global ice volume. However, good reproducibility of benthic $\delta^{18}\text{O}$ records across the global oceans suggests random variability in individuals is unlikely [Lisiecki and Raymo, 2005]. Furthermore, five of the seven sampled intervals exhibit no statistically significant variability among foraminifer tests. In contrast, in the high variability sample windows 10 to 20% of the individuals lie outside of the distributions of the standard or quiescent 200 year interval measurements (Figure 2, 3).

Secondly, anomalous values could also result from sediment mixing via downslope reworking or bioturbation. However, the sedimentology across high variability intervals does not support a role for either process as there is no observable evidence for turbidity flows or burrows from x-radiograph scans of the cores (Figure 4).

Carbon isotopes measured on anomalous individuals are not systematically offset from the rest of the population, which further suggests no downslope movement of foraminifers (Supplemental Information, Text S2; Figure S6, S7). Bioturbation is most likely to introduce spurious variability within low sedimentation rate intervals, but again we observe no correlation between sedimentation rate and foraminifer $\delta^{18}\text{O}$ variability. Finally, the relatively high sedimentation rate, approximately 100 cm/kyrs, throughout the core inhibits bioturbation impacts; the anomalously heavy $\delta^{18}\text{O}$ values observed 7003 yrs BP require organisms to mix deglacial foraminifers up thru more than 300 cm of sediment.

The third and most likely explanation is that outlying $\delta^{18}\text{O}$ values record episodes of anomalously warm or cold temperatures or influxes of fresh or saline water at 1000 m water depth. Interpreted under this framework, our data indicate greater temperature and salinity variability at 1000 m depth in the EEP 4080 yrs BP and 7003 yrs BP relative to today (Supplemental Information, Figure S4). $\delta^{18}\text{O}$ fluctuations must exceed $\sim 0.5\text{‰}$ to give an attribution of variability by our method. Because our time intervals are short and sea level has remained relatively stable during the Holocene, we discount the effects of changing global ice volume on the $\delta^{18}\text{O}$ signature of water and suggest that change in shell $\delta^{18}\text{O}$ dominantly reflects temperature, with a 0.25‰ change in $\delta^{18}\text{O}$ corresponding to $\sim 1^\circ\text{C}$ change in EqPIW [Shackleton, 1974; Fontaneir *et al.*, 2006]. Thus, when attributed fully to temperature, variability within these two intervals provides evidence for temperature variations exceeding 2°C that last for a month or longer. These data therefore imply that modern warming at the site near 1000 m water depth (estimated to be $>0.1^\circ\text{C}/\text{decade}$ [Purkey and Johnson, 2010], or 2°C over 200 yr) is not unprecedented and

falls well within the range of natural variability.

6. Sources of Past and Present Deep Water Variability

Periods of high natural deep ocean variability during the Holocene support strong, rapid transport of signals, whether natural or anthropogenic, by advection, wave or other mechanism, from the surface ocean to at least 1000 m water depth in the EEP. Our data thus provide evidence that the deep ocean is *not* isolated from surface forcing on short timescales, or that significantly more vigorous modes of internal variability are occasionally active. Although the mechanism responsible remains uncertain, our data help constrain the timescale (months to decades) and the magnitude ($>2^{\circ}\text{C}$) of this deep ocean variability, which in turn limits the possible mechanisms. It is difficult to imagine that a shift of local watermasses would be sufficient to explain this signal. In the modern ocean (based on the World Ocean Atlas version 2, *Locarnini et al.*, 2013), seawater with a mean temperature 2°C (roughly 0.5‰ $\delta^{18}\text{O}$ equivalent) or more warmer at 1000 m depth occurs only in the Red Sea, Mediterranean Sea, and the outflows from these seas into the Eastern Indian Ocean and North Atlantic Ocean. Seawater with a mean temperature 2°C or more colder at 1000 m depth occurs only in the Southern Ocean south of 40°S , the Arctic Ocean, or the Bering, Japan, or Okhotsk Seas. Salinity deviations (based on the World Ocean Atlas version 2, *Zweng et al.* 2013) of 2 ppt (rough 0.5‰ $\delta^{18}\text{O}$ equivalent) at this depth occur only in the Black Sea (fresher) or the Mediterranean and Red Seas (saltier). Significant changes to the global circulation can yield this magnitude of variations [*Toggweiler et al.*, 1991; *Kalansky et al.*, 2015; *Bova et al.*, 2015], but not on such short timescales (centennial and faster). It is perhaps easier to imagine a vertical relocation of watermasses, but 2°C warmer water occurs near 660 m

depth, and 2°C colder water occurs near 1750m depth (indicated by the grids in Figure 1). Heaving density surfaces this distance would require internal waves nearly double the world's largest (*Alford et al.*, 2015).

Additional processes linked to deep water variability include: advection, the El Niño Southern Oscillation (ENSO) [*Meehl et al.*, 2011; *England et al.*, 2014], planetary waves and related instabilities [*Johnson et al.*, 2007; *Kawano et al.*, 2006], and mesoscale eddies [*Chaigneau and Pizarro*, 2005; *Chaigneau et al.*, 2011; *Le Bars et al.*, 2016]. However, the timescale and magnitude of intermediate and deep water variability forced by most of these mechanisms are also difficult to reconcile with modern and past observations. The advective timescale is too long [*Masuda et al.*, 2010] and anomalies forced by planetary and internal waves are typically too small (<1°C) [*Nakano and Sugimotohara*, 2002]. Coastal variability associated with El Niño [*Kessler*, 2006], upwelling anomalies [*Huyer et al.*, 1990], and their propagation as Kelvin waves [*Brink et al.*, 1983] are of sufficient magnitude, but this variability occurs primarily within the shallowest 500 m in modern times. Numerous strong ENSO events and shifts in the PDO over the last six decades fail to produce sufficient anomalies at 1000 m depth (Figure 1). Over the past 62 years, the temperature at our study location varies by less than 1°C or just 0.25‰ when converted to shell carbonate $\delta^{18}\text{O}$ (95°W-85°W, 0°-10°S, 1000-1050 m depth, Supplemental Information, Figure S4).

Finally, though short-lived (<120 days), mesoscale eddies are abundant from the coast to about 600-800 km offshore [*Hormazabal et al.*, 2004] and influence subsurface temperature and salinity by roughly the magnitude observed near Peru, but farther to the south [*Chaigneau and Pizarro*, 2005; *Chaigneau et al.*, 2011; *Stramma et al.*, 2013;

Colas et al., 2012]. Perhaps the anomalies observed here represent a relocation of such eddies to our core site. High-resolution modeling of seawater $\delta^{18}\text{O}$ demonstrates that eddies can produce anomalies up to 0.25‰ that persist for months, but does not indicate why some centuries exhibit this variability and others do not at our location [*Stevenson et al.*, 2015]. At 500 m water depth mesoscale eddy activity along the coast of Chile is linked to the ENSO cycle; eddy transport is strongest during La Niña or normal years when horizontal density gradients are at a maximum [*Hormazabal et al.*, 2004]. Although quite a bit shallower in the water column than our study site, it is nevertheless interesting that maximum variability at 1000 m depth in the EEP is documented when ENSO activity was at a minimum at the surface [*Koutavas et al.*, 2012; *Carré et al.*, 2014]. Deep isotope measurements along the coast of Chile and in other regions of strong eddy heat and salt transport, such as the Antarctic Circumpolar Current or the Kuroshio and Gulf Stream extensions may help assess eddy-driven deep ocean variability.

7. Conclusions

Detection of high variability intervals in the Holocene suggests the modern ocean is not inherently more variable than the pre-anthropogenic one. We find two 200-yr intervals during the Holocene with temperature swings greater than about 2°C (~0.5‰ in $\delta^{18}\text{O}_{\text{foram}}$ space) that persist for a month or longer at intermediate depths in the EEP. Whether the detected variability in the temperature of EEP intermediate and deep water masses is merely a local phenomenon, reflecting only mesoscale flow over the rough local bathymetry, or more widespread, affecting regions far afield across both intermediate and deep water depths, remains unknown. Systematic studies of individual benthic foraminifera variability at locations throughout the global oceans will constrain

the geographic extent of variability and provide better statistics of deep ocean variability on sub-centennial timescales that may lead to an enhanced ability to understand, model and predict the role of the deep and intermediate ocean in future global change. Given significant, and widespread variability of the deep ocean on human timescales, new estimates of the global energy budget may include a more important deep-ocean reservoir for excess atmospheric CO₂ and heat with potential consequences for surface temperatures over the coming decades.

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Figures:

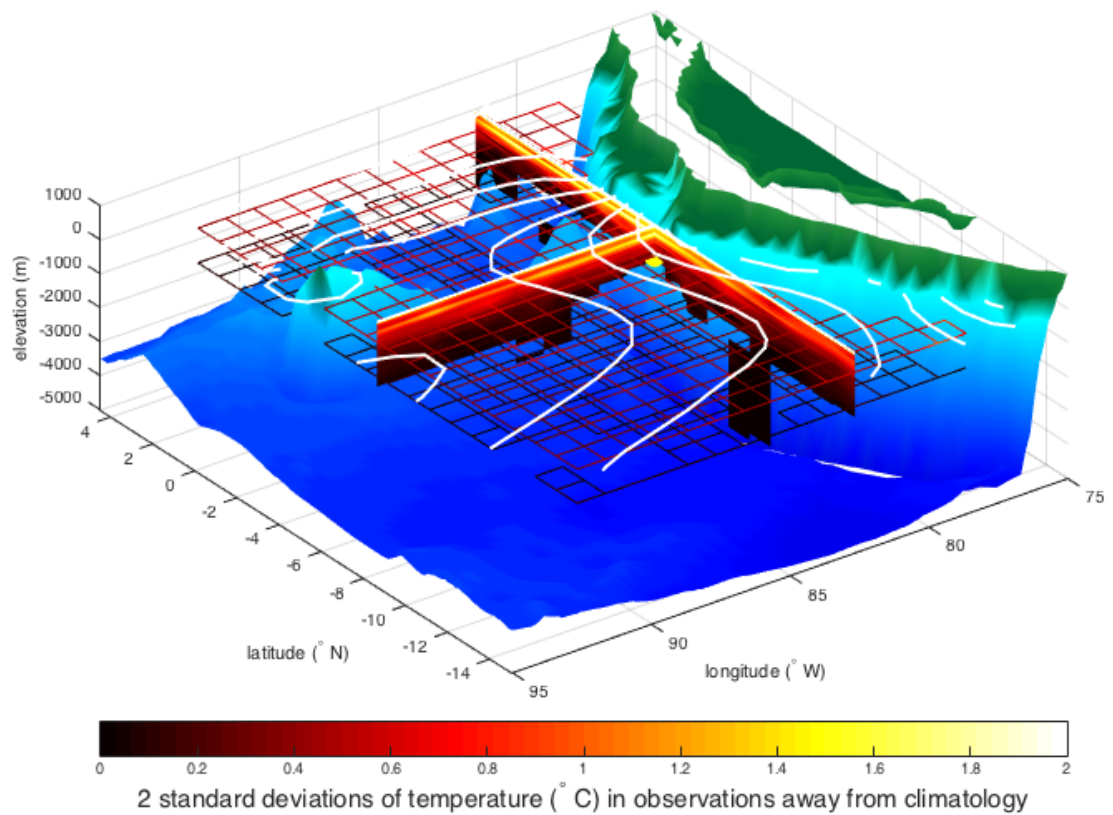


Figure 1. Modern observations of temperature and variability near the CDH-26 and MC-18A core site (yellow ball). Ocean mean surface temperature (white contours, 1°C intervals) and twice the standard deviation away from climatology of all observations (i.e., 95% confidence distance from mean) in °C (shading of curtains and mesh); on this scale isotopic variability detectable using individual foraminifers corresponds to values greater than 1°C, while null episodes are <1°C. The meshes lie on surfaces of potential temperature 2°C above, at 660 m, and below, at 1750 m, the mean temperature at the core site, indicating distance in depth required to achieve a 2°C warming or cooling. Details in the Supplemental Information (Text S2).

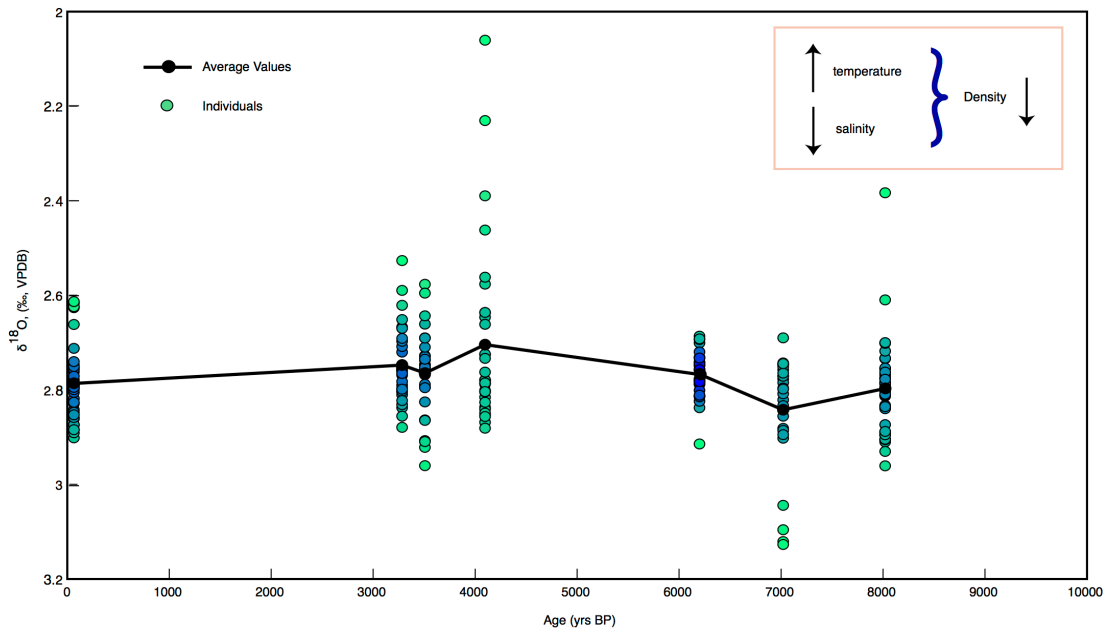


Figure 2. Benthic foraminiferal oxygen isotopes measured on *Uvigerina peregrina*. The mean $\delta^{18}\text{O}$ for each 200 year interval is plotted in black. Individual $\delta^{18}\text{O}$ data are plotted in blue/green with the highest density region of points colored blue.

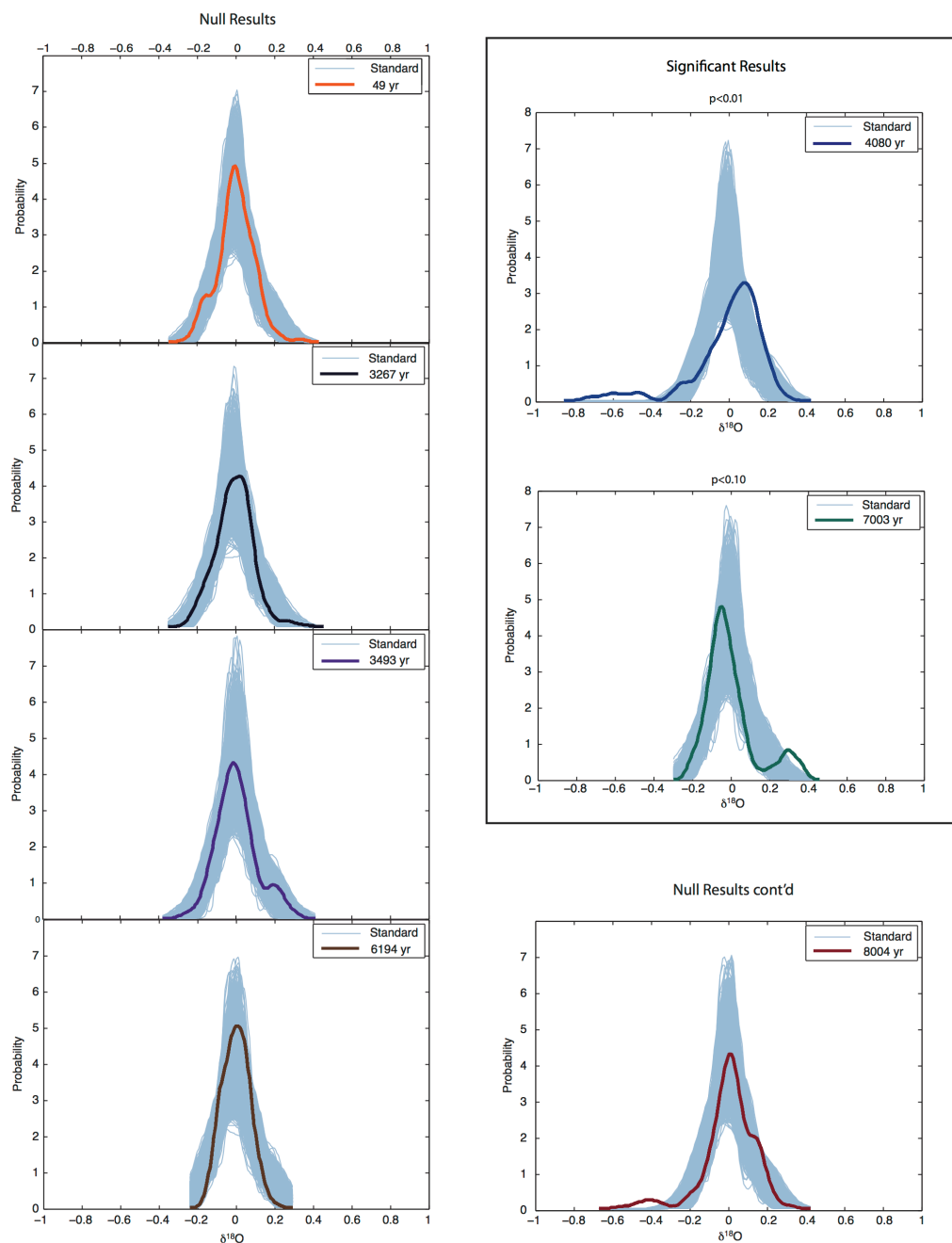


Figure 3. $\delta^{18}\text{O}$ distributions about the mean for each sample interval plotted with 5000 randomly sampled standard distributions (light blue). Distributions from 49, 3267, 3493, 4080, 6194, 7003, and 8004 yrs before present are not distinguishable from the standard distributions. Samples from 4080 and 7003 yrs BP have distributions that have larger

variability than the standards. In the 4080 yr interval the anomaly is a tail consisting of an excess of extreme light isotopic values when compared to the standard. An excess of heavy isotopic values distinguish variability during the interval centered around 7003 yrs BP from the standard.

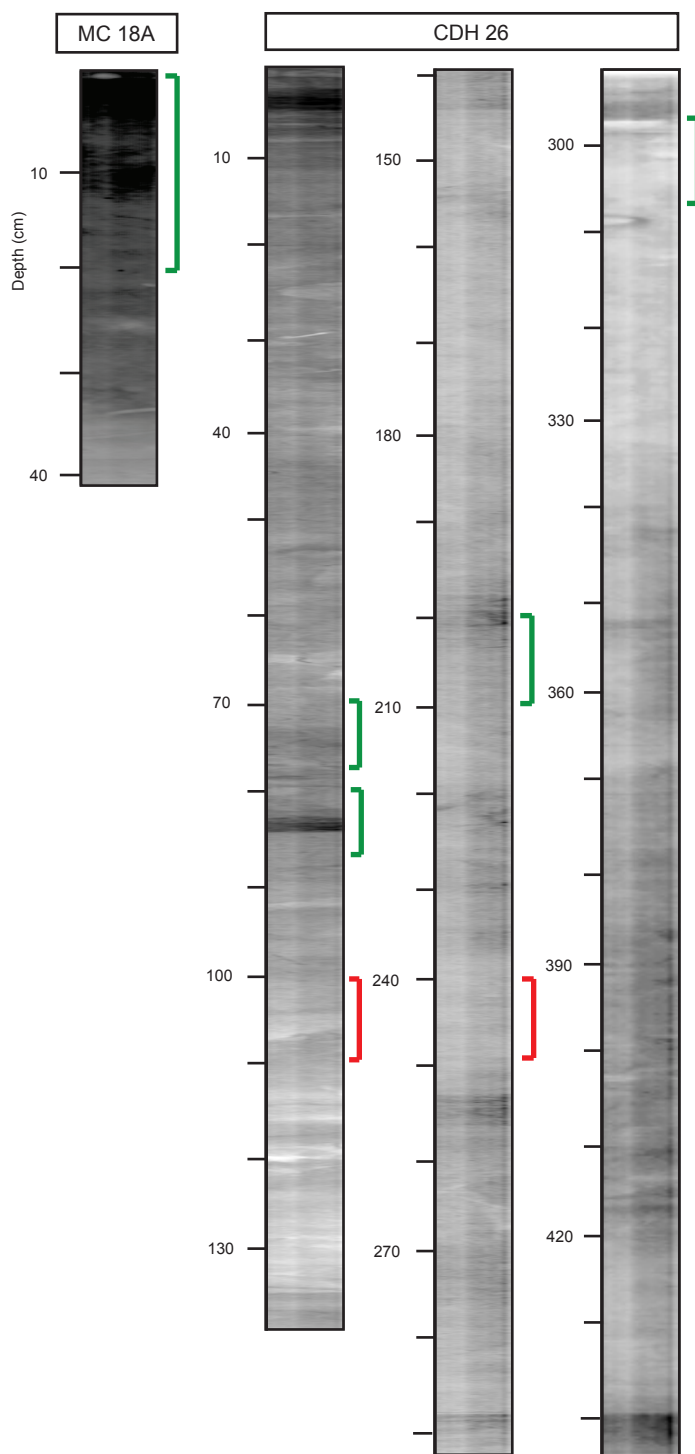


Figure 4. X-radiograph images of core MC-18A and CDH-26. Cores CDH-26 and MC18A were scanned on an ITRAX XRF Core Scanner at the UMASS X-Ray

Fluorescence Laboratory. Light grey indicates low density intervals while the dark grey intervals represent comparatively high density regions. High density intervals tend to have high sand content and may be formed by turbidite flows. Scans were completed at 1 mm resolution. Images were imported into Adobe Photoshop to visually enhance the grayscale density differences throughout the cores.

Table 1. Age range and error for each sampled interval

Core	Core Depth Range (cm)	Upper Age (yrs BP)	Error (1 sd)	Lower Age (yrs BP)	Error (1 sd)	Age span (Best estimate, yrs)	Mean Age (yrs BP)
MC-18A	0 – 20	-56	2.6	153	25.6	208	49
CDH-26	70-76	3190	182	3343	175	153	3267
CDH-26	78.5-86.5	3406	172	3580	165	174	3493
CDH-26	103-110	4008	148	4151	142	143	4080
CDH-26	198-208	6104	76	6284	70	180	6194
CDH-26	239.5-250.5	6910	52	7095	48	185	7003
CDH-26	296-308	7912	36	8095	36	183	8004

* Years BP is referenced to the year 1950. Negative values therefore indicate years after 1950.

Table 2. Variance and range of foraminiferal $\delta^{18}\text{O}$ distributions during each of the seven 200 year intervals sampled and results of K-S tests indicating whether each interval distribution is distinguishable from the standard distribution with and without taking sample sizes into account.

Interval (yrs BP)	Number of Individuals	Std. Deviation (%)	Range (‰)	K-S Test (result, confidence level) all standards	K-S Test Result (result, confidence level) bootstrapped and size weighted standards
-56 – 153	38	0.076	0.41	Null	Null
3190 – 3343	35	0.073	0.47	Null	Null
3406 – 3580	29	0.096	0.43	Null	Null
4008 – 4151	33	0.19	0.95	Distinguishable (99%)	Distinguishable (99%)
6104 – 6384	33	0.053	0.22	Null	Null
6910 – 7095	29	0.11	0.49	Distinguishable (90%)	Distinguishable (90%)
7912 – 8095	37	0.11	0.74	Null	Null
Standard	83	0.010	0.45		

Supplementary Information

Text S1. Data Quality Controls

Individual benthic foraminifera can be difficult to measure to the same level of precision as conventionally sized samples because they are smaller by mass and therefore produce less CO₂. Problems arise because ¹⁸O is present at much lower concentrations relative to ¹⁶O and can therefore be difficult to measure accurately for small samples. As a result, mass spectrometers may exhibit a “linearity effect”, essentially a drift with sample size. Exceedingly small samples may also produce so little CO₂ that gas flow in the capillary lines transitions from laminar to turbulent flow, resulting in isotopic fractionation. At small sample sizes it is difficult to balance the sample and reference gas pressures. During analyses of small samples the sample gas pressure was systematically lower than that of the standard, leading to a systematic offset towards light δ¹⁸O values when the sample gas signal fell below 800 mV on cycle 1 (Fig. S3). Determining the sample size transition point into turbulent flow and the linearity of your instrument is therefore important, particularly when the target signal is small. Prior to making any individual benthic measurements we therefore ran 102 standards ranging from 4 to 111 µg of calcium carbonate, each with a known oxygen isotopic value of -6.52 ± 0.06 ‰ to detect these biases. Results from this test are shown in Figure S3.

Controlling for Turbulent Flow

Symptoms of turbulent flow are detected in measurements made on less than ~800 mV of mass 44 CO₂ (Supplementary Figure 3). We therefore eliminate all measurements made below this threshold. All samples analyzed on the IR-MS are measured 8 times. For a conventionally sized sample, all 8 cycles are averaged to

minimize random errors. For small samples, however, it may be advantageous to remove some measurement cycles because the abundance of mass 44 CO₂ decreases with each cycle. In some cases, the initial measurement cycles are made when the mass 44 CO₂ signal is greater than 800 mV and later measurements on less. As long as 5 of the 8 cycles have a CO₂ mass 44 signal greater than 800 mV we remove just the cycles less than 800 mV and keep the remainder in the dataset. When more than 5 of the 8 cycles measured are less than 800 mV, however, we eliminate the sample entirely.

Controlling for Random Errors

After controlling for variability induced by turbulent flow in the IR-MS, we evaluate the data for random error during analysis. We control for random error by assessing the standard deviation of $\delta^{18}\text{O}$ values measured across one sample's 8 measurement cycles (or as few as 5 if we removed some of the cycles). We remove samples whose $\delta^{18}\text{O}$ values exhibit a standard deviation greater than 0.08‰ from the dataset.

Controlling for Linearity Effects

A small linearity effect is detected in samples that produce between 800 and 2000 mV of mass 44 CO₂, with smaller samples in this range shifted systematically towards heavier $\delta^{18}\text{O}$ values. We do not explicitly correct the individual data for linearity effects when comparing variability between foraminiferal $\delta^{18}\text{O}$ assemblages and the small standard distributions (Figure 3). Instead, K-S tests are performed between sample and standard $\delta^{18}\text{O}$ distributions only if they are indistinguishable based on size. First, we randomly select a 29-37 member set of the 81 small standards with replacement to match the sample being considered in number (e.g. the 3300 yr sample of 34 individual

foraminifer measurements is compared to 34 standard measurements). A K-S test between the sample and standard size distributions must be passed at the 95% confidence level to proceed. If so, a second K-S test is performed to compare the standard and sample $\delta^{18}\text{O}$ distributions (Figure 3). This procedure was repeated 5000 times.

Text S2. Supplementary Results

*Carbon Isotopic Variability of Individual *U. peregrina**

U. peregrina live in the sediment, inhabiting the top 2 cm of the sediment. Studies show *U. peregrina* calcify in approximate equilibrium with the overlying seawater and provide reliable records of seawater $\delta^{18}\text{O}$ and temperature through time [e.g. Shackleton, 1974], but their $\delta^{13}\text{C}$ signature is prone to overprinting by porewater processes (i.e. respiration) [McCorkle *et al.*, 1990, 1997; Fontanier *et al.*, 2006]. To evaluate the impact of porewater chemistry on shell stable isotope signatures we include the carbon isotopic values from the same shells used for oxygen isotope analysis here (Supplementary Figures 7 and 8). Carbon isotopic variance and oxygen isotopic variance have no significant correlation, and shells with anomalous $\delta^{18}\text{O}$ values were not more likely to have anomalous $\delta^{13}\text{C}$ values. All intervals exhibit $\delta^{13}\text{C}$ distributions that are more variable than the lab standards (Supplementary Figure 8).

Modern Observations

The modern observations used to create Figure 1 and Supplementary Figure S5 were derived from the averaged decadal averages from 1955 to 2012 of the World Ocean Atlas [Locarnini *et al.*, 2013] and all high-quality in situ temperature observations from the World Ocean Database [Boyer *et al.*, 2013] respectively. The near-surface temperature and depth corresponding to surface-referenced potential temperatures 2°C

above (6.629°C) and 2°C below (2.629°C) were derived from the average of the decadal averages of the World Ocean Atlas. The standard deviations used for shading in Figure 1 are the standard deviations from all observations at each site away from the climatology.

Supplementary Figure 5 is formed from binning all World Ocean Database observations within 25 m of the target depth from the National Ocean Database Center collection of hydrographic stations (n=177,152 observations at 0-50m, n=152,791 at 225-275m, n=154,689 at 475-525m, n=25,296 at 1000-1050m, and n=23,875 at 1200-1250m). Directly observed variations of surface values in $\delta^{18}\text{O}$ from the *LeGrande & Schmidt* (2006) database (<http://data.giss.nasa.gov/o18data>) are similarly spread in magnitude to the measured spread of 100 yr sediment foraminifera ensembles. In this location, only surface observations of $\delta^{18}\text{O}$ have been collected.

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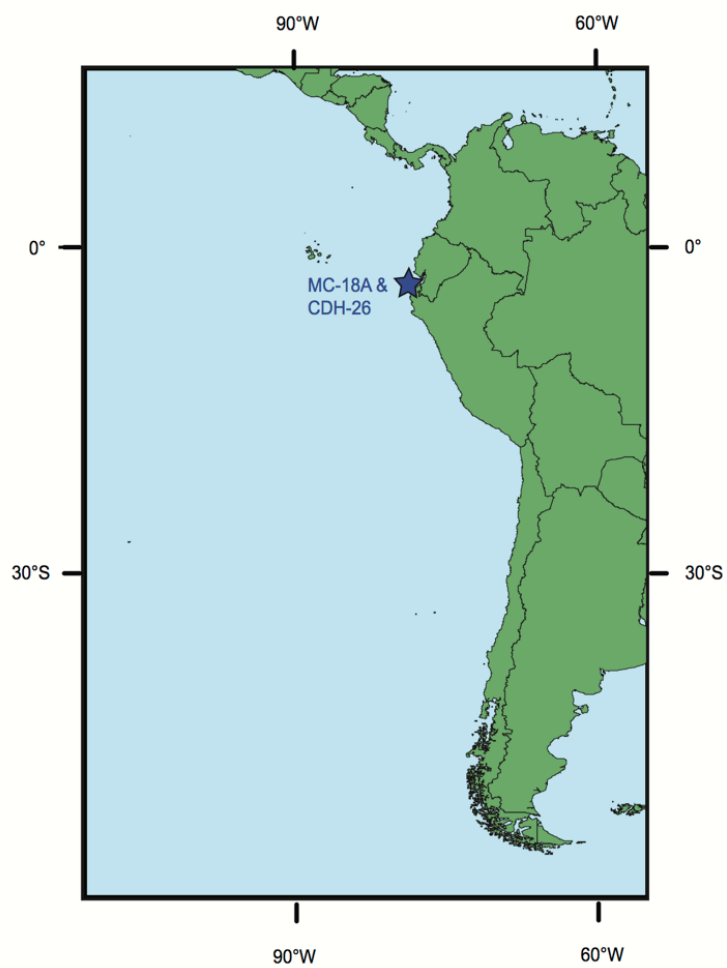


Figure S1. Site map showing locations of the sediment cores, CDH-26 and MC-18A, used in this study.

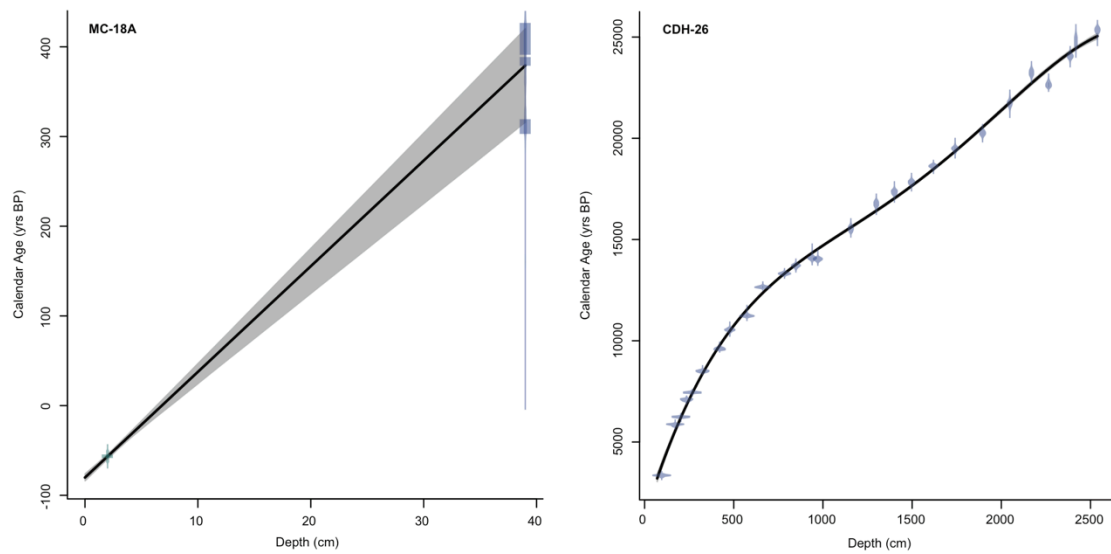


Figure S2. Age model for core CDH-26 and MC-18A based on 27 and 2 AMS ^{14}C dates, respectively.

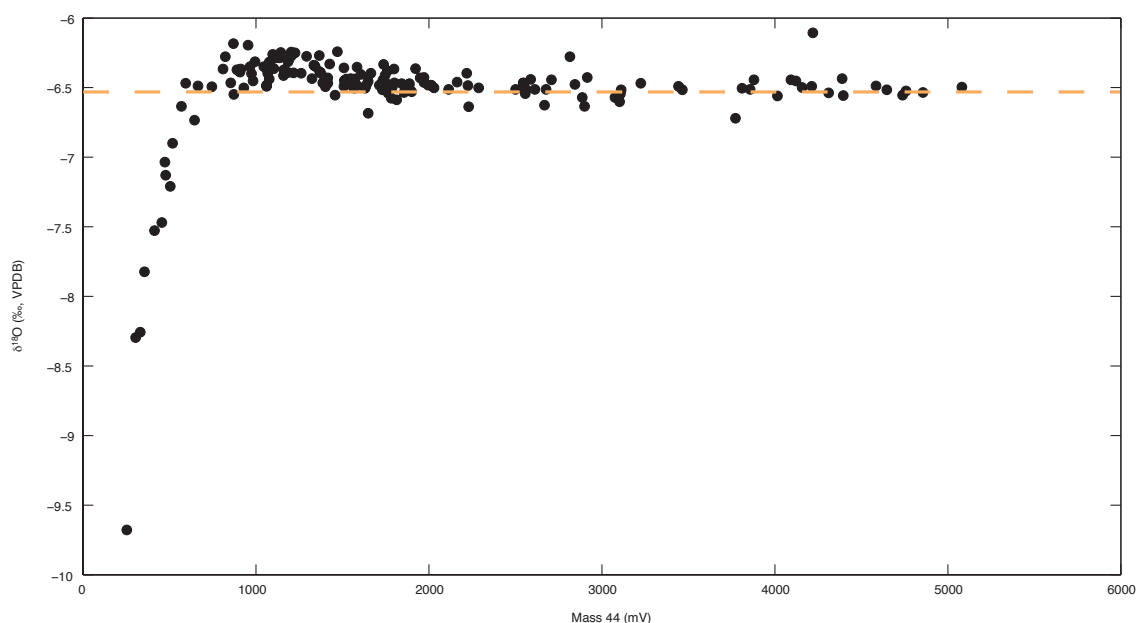
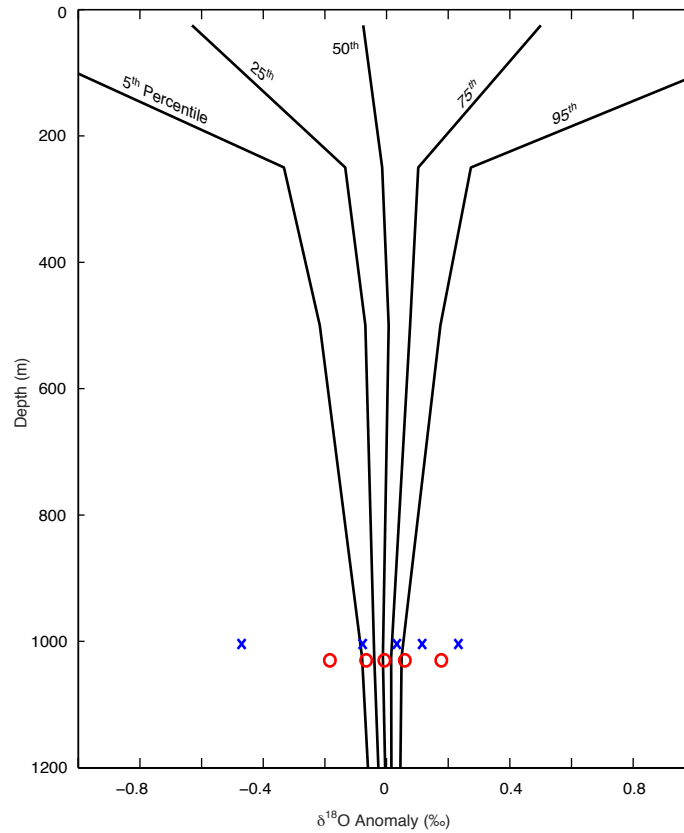


Figure S3. Oxygen isotopes of a lab standard as a function of the mass 44 CO₂ voltage signature on cycle one of eight. Sample size increases with mass 44 CO₂ voltage signals. The orange dashed lines indicates the true standard δ¹⁸O signature of -6.52‰. Between 800 and 3000 mV, δ¹⁸O signatures trend towards anomalously heavy values with decreasing mass 44 cycle 1 voltages as a result of linearity effects, which leads to an increase in δ¹⁸O signatures decrease by 0.06‰ per volt. Below 800 mV on cycle 1, the sample and reference gas pressures are not balanced and the sample gas transitions to turbulent flow before the reference gas. As a result, δ¹⁸O signatures become increasingly light at low mass 44 voltages. To avoid biases resulting from poor balance between the sample and reference gas at low sample size we eliminate any measurement on <800 mV at the mass 44 detector on cycle 1.



Supplementary S4. Modern observations [all dates up to 2012, from *Boyer et al.*, 2013] located between 95°W and 85°W and 10°S and 0° and 100 m vertical bins are used to categorize variability in this region. Solid lines indicate the depth dependence in estimated $\delta^{18}\text{O}$ deviation from the mean based on observed temperatures. Blue crosses and red circles indicate the same percentiles from the 4080 yr and 49 yr sediment foraminifera ensembles, respectively.

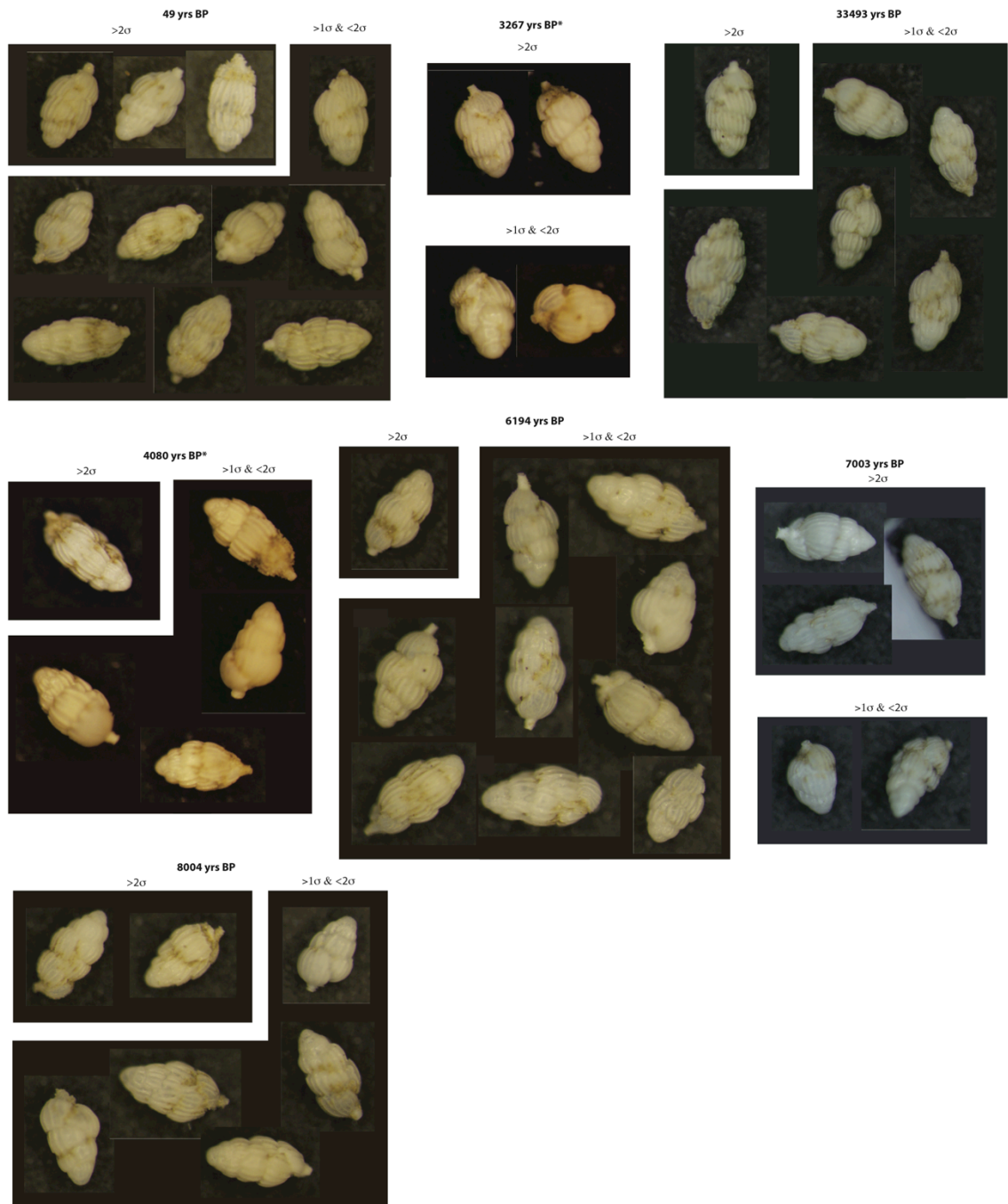


Figure S5. Images of individual foraminifera whose measured $\delta^{18}\text{O}$ signatures were $\geq \pm 1\sigma$ from the mean $\delta^{18}\text{O}$ value calculated for each sampling interval. *Images are unavailable for 1 foraminifer with a measured $\delta^{18}\text{O}$ value more than 2σ from the mean $\delta^{18}\text{O}$ value 4080 years BP and 5 images are unavailable for foraminifera more than 1σ

from the mean $\delta^{18}\text{O}$ value 3267 years BP. These intervals were the first to be measured and we only began photographing the foraminifers after the first round of results.

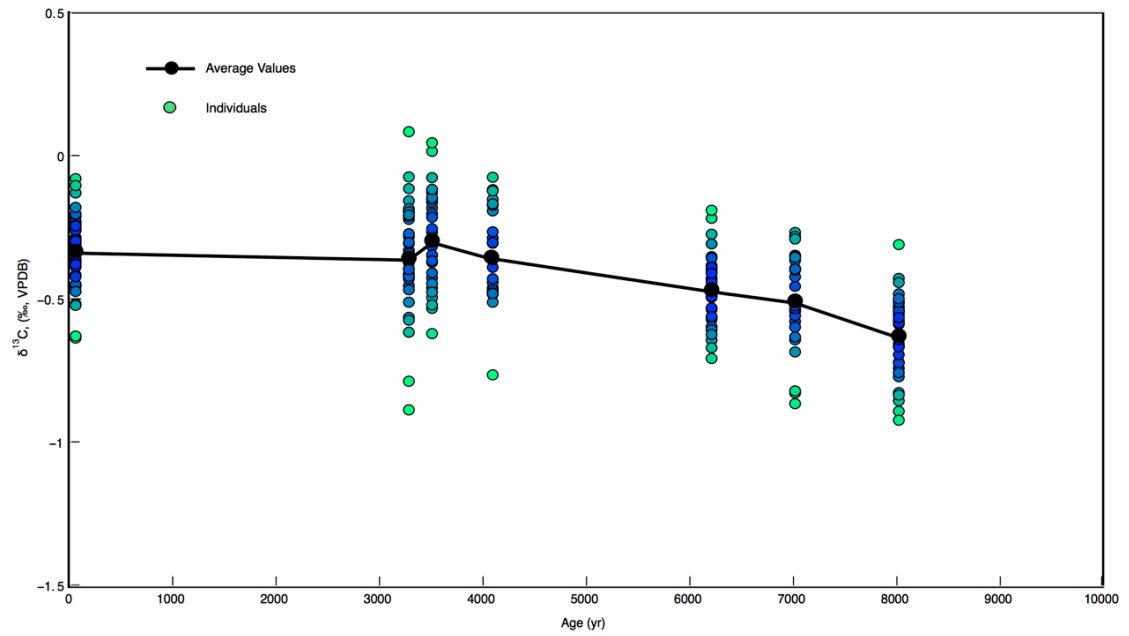


Figure S6. Benthic foraminiferal carbon isotopic values measured on *Uvigerina peregrina*. The mean $\delta^{13}\text{C}$ for each ~200 year interval is plotted in black. Individual $\delta^{13}\text{C}$ data is plotted in blue/green with the highest density region of points colored blue.

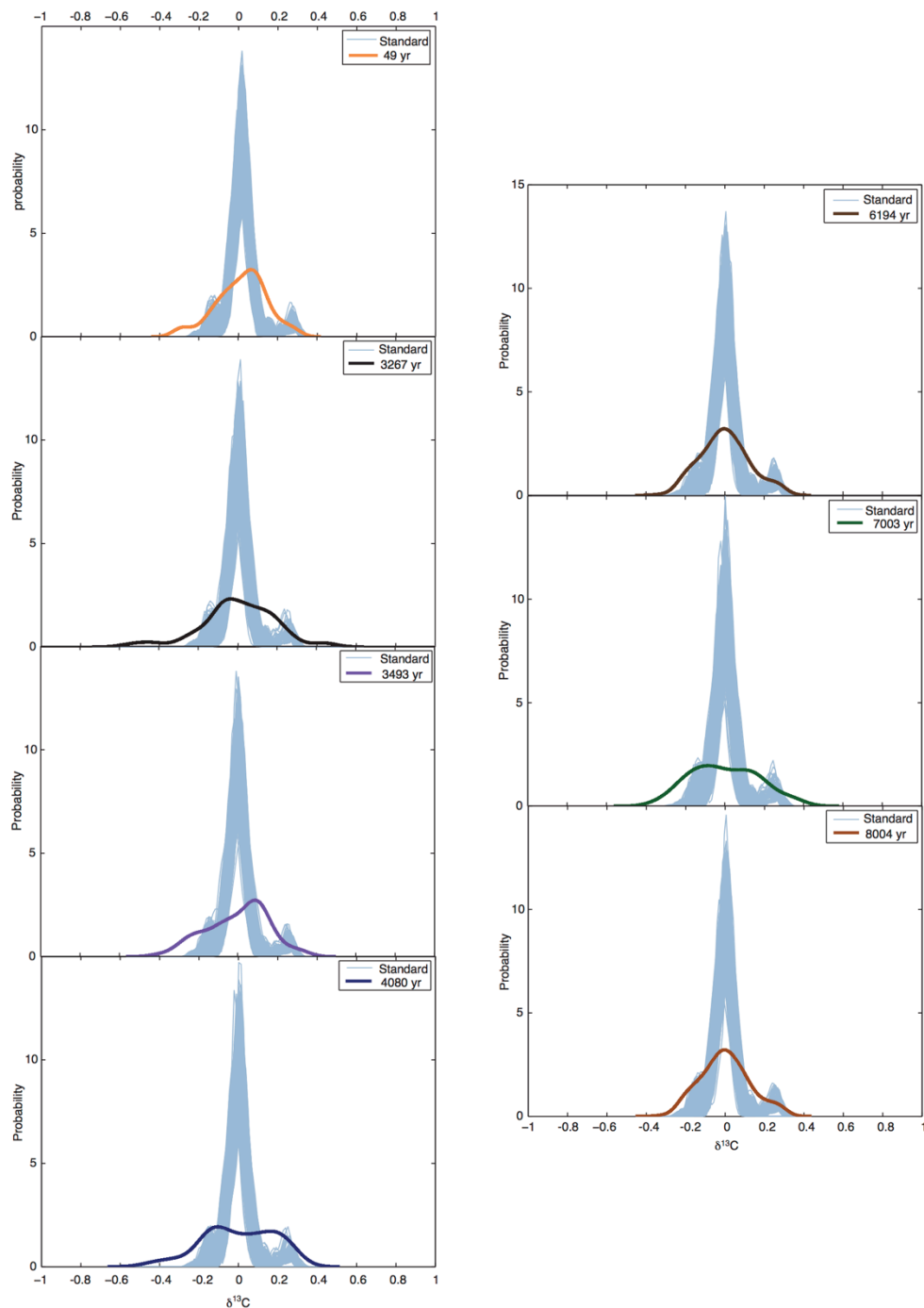


Figure S7. $\delta^{13}\text{C}$ distributions about the mean for each sample interval plotted with 5000 randomly sampled standard distributions (light blue). Distributions from all intervals are distinguishable from the standard distributions at high confidence, likely due to

overprinting by porewater processes that affect shell $\delta^{13}\text{C}$ but not $\delta^{18}\text{O}$ signatures

[McCorkle *et al.*, 1990, 1997; Fontanier *et al.*, 2006].

CHAPTER THREE

Radiocarbon evidence for storage and release of the glacial carbon reservoir in the equatorial Pacific

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Abstract

Atmospheric carbon dioxide rose by ~80 ppmv during the last deglaciation, which helped propel Earth's climate into the present interglacial. Identifying the source and release pathway of this carbon is critical for understanding glacial-interglacial feedbacks within the climate system. The prevailing hypothesis suggests carbon accumulated in the deep Southern Ocean, becoming increasingly ^{14}C -depleted during the last glacial period, and was released into southern ocean intermediate waters (SOIWs) and, subsequently, the atmosphere during deglaciation. In this paper, we present new radiocarbon records from three intermediate depths in the Eastern Equatorial Pacific that challenge aspects of this paradigm. We find that SOIWs were *not* the primary conduit for transporting upwelled old carbon from the Southern Ocean to the atmosphere. Instead, aged carbon was isolated in both the deep north and deep south Pacific and injected directly into the intermediate depth equatorial Pacific.

1. Introduction

Atmospheric carbon dioxide (CO_2) levels are an excellent indicator of Earth's climate state. Ice core records from Antarctica demonstrate a tight coupling between global temperatures and atmospheric CO_2 over the past 420,000 years, with shifts in atmospheric CO_2 accounting for close to half of the radiative forcing required to move between glacial and interglacial states [Weaver *et al.*, 1998; Yoshimori *et al.*, 2001]. Since these changes were first observed nearly 40 years ago [Berner *et al.*, 1979], studies of oceanic radiocarbon have demonstrated that the deep ocean is capable of storing and releasing enough carbon to explain glacial-interglacial CO_2 changes [Sarnthein *et al.*, 2013; Skinner *et al.*, 2015]. The requirements for carbon storage within the deep ocean

are twofold: (1) an influx of organic matter from the surface ocean and (2) isolation of deep ocean waters from the surface ocean to prevent gas exchange with the atmosphere. Evidence for these conditions should be evident in the proxy record as large chemical gradients between surface and deep waters and the build-up of old, ^{14}C -depleted, nutrient-rich, and oxygen poor waters in the subsurface.

Aged deep water masses have been detected during the Last Glacial Maximum (LGM) at numerous sites throughout the global ocean [Keigwin, 2004; Robinson *et al.*, 2005; Galbraith *et al.*, 2007; Skinner *et al.*, 2004; 2010; 2014; 2015; Lund *et al.*, 2011; Okazaki *et al.*, 2012; Cook and Keigwin, 2015; Ronge *et al.*, 2016; Sikes *et al.*, 2016]. LGM radiocarbon signatures from intermediate and deep water masses in the southwestern and south central Pacific provide evidence for aged carbon within Pacific Deep Water (PDW) [Ronge *et al.*, 2016]. Radiocarbon signatures within PDW in the North Pacific are depleted as well, suggesting widespread reduction in the ventilation of PDW relative to today [Galbraith *et al.*, 2007; Lund *et al.*, 2011; Cook and Keigwin, 2015]. Occurrences of old glacial deep waters are also observed in the South Atlantic and North Atlantic [Skinner *et al.*, 2004; 2010; 2014], with moderately aged deep water in the Drake Passage [Burke and Robinson, 2012; Chen *et al.*, 2015]. Although the magnitude of the ^{14}C anomaly in the deep ocean is not consistent at all sites, glacial water mass ventilation ages are consistently older at depth in each basin than those observed today, supporting the presence of a deep glacial carbon reservoir.

During the last deglaciation, the aged deep carbon pool disappeared, likely fueling the two-part rise in atmospheric CO_2 [Monnin *et al.*, 2004; Skinner *et al.*, 2010; 2014; Marcott *et al.*, 2014; Ronge *et al.*, 2016; Sikes *et al.*, 2016]. Enhanced upwelling around

the Antarctic continent triggered by poleward shifts in the position of the westerly winds, reduced sea-ice extent, salinity driven buoyancy differences in Southern Ocean, or some combination of the three [e.g. *Keeling et al.*, 2001; *Toggweiler et al.*, 2006; *Watson et al.*, 2015] reinvigorated deep ocean circulation, potentially ventilating aged, ^{14}C -depleted carbon to the Southern Ocean surface. Because carbon equilibration with the atmosphere is slow, on the order of a year, the majority of upwelled carbon would have been re-sequestered in Antarctic Intermediate Water (AAIW) and spread throughout the shallow sub-surface oceans.

Strongly depleted ^{14}C -concentrations have been documented at four locations bathed by AAIW coincident with atmospheric CO_2 rise during deglaciation, suggesting the water mass was an important conduit for the release of CO_2 to the atmosphere [*Marchitto et al.*, 2007; *Lindsey et al.*, 2015; *Bryan et al.*, 2010; *Stott et al.*, 2009; *Basak et al.*, 2010]. Efforts to corroborate these findings at sites close to the Antarctic Intermediate Water formation region, however, find little to no change in the water mass ventilation ages, with all those that exhibit signatures of old carbon located at the furthest reaches of AAIW, within the tropical Pacific and Indian Oceans (Fig. 1) [*Kennett and Ingram*, 1995; *Marchitto et al.*, 2007; *Stott et al.*, 2009; *Bryan et al.*, 2010; *De Pol-Holz et al.*, 2010; 2012; *Magini et al.*, 2010; *Rose and Sikes*, 2010; *Sorter et al.*, 2011; *Burke and Robinson*, 2012; *Siani et al.*, 2013; *Romahn et al.*, 2014]. If AAIW was the conduit for old carbon the most ^{14}C -depleted signatures should be observed closest to the Southern Ocean source region.

In this study, we evaluate the radiocarbon history of intermediate and surface water masses in the eastern equatorial Pacific over the past 25 kyrs. Three new paired

planktonic and benthic foraminiferal radiocarbon records from the Northern Peru Margin, at 373 and 1023 m water depth (CDH 23: 03°44.95' S, 81°08.05' W and CDH 26: 03°59.16' S, 81°18.52' W, respectively), and the Galapagos Platform, at ~600 m water depth (GGC 43: 01°15.13'S, 89°41.07'W, 617 m and CDH41: 01°15.94'S, 89°41.88'W, 595 m), provide a regional profile of water column radiocarbon variability from the tropical Pacific. We find that the radiocarbon signature of waters at 600 m depth on the Galapagos platform were strongly influenced by the deep Pacific carbon reservoir during both the LGM and the deglaciation. During the LGM, deep water ^{14}C -concentrations, as recorded by benthic foraminifera, were strongly depleted and are comparable to signatures typically observed much deeper in the water column, while deglacial ^{14}C trends align with intermediate water radiocarbon records from the Baja Peninsula [Marchitto *et al.*, 2007; Lindsay *et al.*, 2015] and the Arabian Sea [Bryan *et al.*, 2010]. This dataset therefore supports a strong link between the cold, dense waters of the deep ocean and the warm, tropical waters of the Pacific and a new mechanism for the release of the glacial carbon reservoir outside the Southern Ocean.

2. Results

2.1 Core Chronology and Benthic-Planktonic age offsets

We measured a total of fifty-seven paired planktonic and benthic ^{14}C dates between the three EEP sites (Appendix C, Fig. 3). Today, all cores are bathed by Equatorial Pacific Intermediate Water (EqPIW), a mixture of Antarctic Intermediate Water (AAIW) with contributions from Pacific Deep Water (PDW) and Subantarctic Mode Water (Fig. 2) [Bostock *et al.*, 2010; 2013]. Planktonic ^{14}C dates, measured on *Neoglobobulimina dutertrei* (CDH 23 and CDH 26) and *Globobulimina ruber* (GGC

43 and CDH41), were collected as part of a previous study [Bova et al., 2015] and paired here with benthic ^{14}C ages to calculate age offsets between surface and deep water masses. Benthic ^{14}C dates were measured on mixed benthics (*Bolivina spp.*, *Hanzawaia concentrica*, and *Uvigerina spp.*) at the Peru Margin sites and *Uvigerina spp.* at the Galapagos.

Core chronologies were constructed using a polynomial fit to sets of planktonic AMS ^{14}C dates using the *Clam version 2.2* age-modeling software package for R (Supplementary Fig. 1) [Blaauw, 2010]. All ^{14}C dates were converted to calendar ages using the marine13 radiocarbon age calibration curve [Reimer et al., 2013]. We modify all age models from the published chronologies of Bova et al. [2015] in order to characterize age model uncertainty using the *Clam* software. In addition, we remove 7 planktonic ^{14}C dates from the chronology for cores GGC 43 and CDH 41 (617 and 595 m, respectively).

In the Galapagos Platform cores, paired benthic and planktonic foraminiferal ^{14}C ages converge approximately during the Younger Dryas (YD) and Heinrich Stadial 1 (HS1), possibly indicating enhanced mixing of ^{14}C -depleted deep waters to the surface ocean and shifting surface ^{14}C ages to artificially old ages (Fig. 3). We remove planktonic ^{14}C dates from these susceptible intervals (i.e. ~YD and HS1) and assume a constant 500 year reservoir age correction based on modern observations from the marine reservoir age database [Taylor and Berger, 1967; Druffel, 1981; Etayo-Cadavid et al., 2013].

The difference between benthic and planktonic ^{14}C ages (B-P) typically reflects age since water mass ventilation. The B-P ^{14}C ages from the Galapagos Platform indicate larger offsets during the last glacial period than during the Holocene by 2000 to 3500

years (Fig. 4b). B-P offsets decline during the deglaciation in two phases. They first decrease by 2200 years 18.4 ka. Then, after a brief rebound to near glacial values from 14.9 to 12.4 ka, they shift abruptly to just 150 yrs at the end of the deglaciation within 1.5 kyrs. Holocene B-P offsets at 600 m are relatively stable, averaging ~315 years. The records from 370 and 1000 m water depth along the northern Peru margin exhibit markedly different B-P ventilation age histories. At 1000 m water depth B-P ages are largest during the LGM, with an average offset of 1125 yrs, about half that observed between surface waters and water masses at 600 m water depth near the Galapagos, approximately 1000 km to the west. At the end of the deglaciation, B-P offsets decrease to 465 years during the early Holocene. ^{14}C offsets between the surface and 373 m are stable, averaging 260 years over the entirety of the record.

2.2 Benthic Foraminiferal $\Delta^{14}\text{C}$

Benthic foraminiferal $\Delta^{14}\text{C}$ estimates the radiocarbon activity of the water mass in which the individuals grew. We calculate this value from the foraminiferal AMS ^{14}C date and its estimated calendar age using the equation of *Adkins and Boyle* [1997], which accounts for decay since the foraminifera were alive. The calendar age is the largest source of uncertainty in this calculation.

Patterns in $\Delta^{14}\text{C}$ at all water depths are similar, but opposite in sign to trends in B-P, with reduced $\Delta^{14}\text{C}$ values generally corresponding to larger B-P offsets (Fig. 4a). Holocene $\Delta^{14}\text{C}$ values from all water depths track atmospheric $\Delta^{14}\text{C}$ values, exhibiting a near constant 100‰ offset. $\Delta^{14}\text{C}$ signatures at 600 and 1026 m water depth exhibit significantly larger offsets from the atmosphere during the deglaciation and LGM, on average ~402 and 254‰, respectively. The Galapagos record (600 m) exhibits the lowest

$\Delta^{14}\text{C}$ values of the three sites, with minor enrichments occurring 16.8 ka before plunging back towards glacial values during the YD. At the end of the YD, $\Delta^{14}\text{C}$ values at 600 m shift from a 518‰ offset from atmospheric $\Delta^{14}\text{C}$ values to 91‰ offset in just over 1.5 kyrs. $\Delta\Delta^{14}\text{C}$ offsets at 600 and 1000 m water depth are largest during the LGM, shifting to modern values abruptly at the end of the YD interval. The record from 373 m differs from the deeper records in that its $\Delta^{14}\text{C}$ values remain constant during the YD.

3. Discussion

Radiocarbon records from within AAIW are valuable for constraining the mechanism responsible for glacial-interglacial atmospheric CO_2 variability because the water mass is the hypothesized conduit between a deep ocean glacial carbon reservoir and the atmosphere. In this study, we present radiocarbon records of EqPIW, a mixture of predominantly AAIW with contributions from PDW, in the EEP. Two records from the far eastern equatorial Pacific, along the N. Peru Margin, from 370 and 1000 m exhibit little or no evidence for aged waters at the core sites during the LGM or deglaciation. Our $\Delta^{14}\text{C}$ record from 600 m water depth on the Galapagos platform, however, is tightly coupled to intermediate depth $\Delta^{14}\text{C}$ record from off the western coast of Baja California (23.5°N, 111.6°W) during the last deglaciation (Fig. 5) [Marchitto *et al.*, 2007; Lindsay *et al.*, 2015]. Like our sediment cores, EqPIW bathes the Baja sediment cores used in the 2007 study [Bostock *et al.*, 2010]. $\Delta^{14}\text{C}$ records from Baja and the Galapagos are strongly depleted during the last deglaciation, with minimum $\Delta^{14}\text{C}$ values between -200 and -300‰ observed at both sites during the YD interval. During the LGM, however, the records diverge. In fact, our record from the intermediate depth Galapagos region is unlike any other published intermediate depth radiocarbon record, with $\Delta^{14}\text{C}$ values in the

LGM comparable to some of the most ^{14}C -depleted deep water masses documented at the time (Fig. 5).

A synthesis of the presently available radiocarbon records spanning the past 30 kyrs, including those presented here, reveals characteristic radiocarbon histories by water mass in the Pacific Ocean (Fig. 5). Benthic foraminiferal radiocarbon records recovered from AAIW depths maintain a relatively constant ^{14}C offset from the atmosphere over the past 30 kyrs [De Pol-Holz *et al.*, 2010; Rose and Sikes, 2010; Sorter *et al.*, 2011; Burke and Robinson, 2012; Siani *et al.*, 2013]. The Peru Margin records (this study), from 373 and 1026 m, compare well with these records. In contrast, intermediate depth sites bathed by EqPIW, a mixture of primarily AAIW with contributions from PDW, exhibit $\Delta^{14}\text{C}$ depletions of -200 to -300‰ during the last deglaciation [Marchitto *et al.*, 2007; Bryan *et al.*, 2010; Romahn *et al.*, 2014]. North PDW (NPDW) $\Delta^{14}\text{C}$ values were as low as -181‰ and 8‰ during the last deglaciation and LGM, respectively, while the ^{14}C signature of South PDW (SPDW), as revealed in the Atlantic [Skinner *et al.*, 2010] and Pacific [Skinner *et al.*, 2015; Ronge *et al.*, 2016; Sikes *et al.*, 2016] sectors of the Southern Ocean, is still more depleted during the LGM. In fact, ^{14}C concentrations are so low at some south Pacific locations ($\Delta^{14}\text{C}$ reached -600‰ in the south Pacific) that hydrothermal CO_2 inputs of ^{14}C -free carbon to an already isolated reservoir are likely required given the apparent absence of widespread anoxia [Ronge *et al.*, 2016]. Regardless of the ultimate CO_2 source, these records indicate that ^{14}C -depleted carbon accumulated within PDW during the last glacial period and was released into EqPIW during the last deglaciation. The 600 m core site on the Galapagos Platform appears to be uniquely positioned such that the dominant water mass at the core site switched from

PDW during the LGM to EqPIW during deglaciation, recording both the PDW glacial reservoir and its deglacial release.

PDW is the dominant deep water mass in the EEP at depths greater than ~1200 m today but its signature can be traced much shallower in the water column [Bostock *et al.*, 2010]. AAIW is the main component of EEP intermediate waters but the high salinity, nutrient concentrations, alkalinity, and DIC as well as the low oxygen and $\Delta^{14}\text{C}$ signature of these waters require contributions from depth, most likely PDW [Bostock *et al.*, 2010]. Diapycnal mixing of PDW to intermediate depths is not complete and a large $\Delta^{14}\text{C}$ gradient (~100‰) is maintained between the water masses today [Key *et al.*, 2015]. During the LGM, however, $\Delta^{14}\text{C}$ signatures at 600 m water depth near the Galapagos Islands are within error of those observed in the heart of PDW, at 2921 m water depth, at site ODP 1240, and at five other PDW sites from the Atlantic and Pacific sectors of the Southern Ocean [Skinner *et al.*, 2010; de la Fuente *et al.*, 2015; Sikes *et al.*, 2016] and the North Pacific [Galbraith *et al.*, 2007; Lund *et al.*, 2011; Cook and Keigwin, 2015], suggesting enhanced mixing of PDW to at least 600 m water depth near the Galapagos Islands and challenging the local presence of a bathyal front at 2000 m during glacial times [Herguera *et al.*, 1992]. We suggest contributions from aged PDW can explain the sustained presence of a ^{14}C -depleted water mass to at least 600 m on the Galapagos Platform during the LGM.

Penetration of PDW to shallow subsurface depths was not ubiquitous in the EEP. Along the Peru Margin the modern $\Delta^{14}\text{C}$ gradient between intermediate waters and PDW was maintained during the LGM (Fig. 5), which suggests that the aged water mass identified at 600 m on the Galapagos platform was localized to either a limited vertical

range, between 370 and 1000 m, and/or longitudinally constrained west of the continental margin. Water mass ages at 370 and 1000 m water depth along the margin are comparable to modern ^{14}C based ventilation age calculations from the nearest GLODAP station (85.833°W , 1.999°S) over the duration of both records [Key *et al.*, 2015]. The high regional heterogeneity is surprising given the proximity of the core sites but is likely related to the complexity of ocean currents in the region. Today there is no North Pacific source to thermocline waters east of the Galapagos, and a sharp boundary at the equator between South and North Pacific contributions is identified based on tritium and ^3He sections from WOCE [Jenkins, 1996; Fiedler and Talley, 2006]. In addition, interaction of currents with the shallow, rough bathymetry surrounding the Galapagos Islands may enhance mixing, and can induce leeside formation of eddies, fronts, and localized upwelling possibly increasing the influence of deep water masses, specifically PDW, at relatively shallow depths in the vicinity of the islands [Heywood *et al.*, 1990; Caldeira *et al.*, 2002].

We therefore suggest that pulses of old carbon observed at intermediate depths near the Galapagos, Baja, and Oman coasts during the last deglaciation represent the release of aged, ^{14}C -depleted carbon from PDW. Strongly reduced $\Delta^{14}\text{C}$ signatures characteristic of PDW in the Atlantic and Pacific sectors of the Southern Ocean during the LGM increase beginning 18 ka and reach modern values by the end of HS1 [Skinner *et al.*, 2010; 2015; Sikes *et al.*, 2016; Ronge *et al.*, 2016]. Some of the exceptionally aged carbon stored within PDW in the Southern Ocean may have escaped via Southern Ocean upwelling and transport by AAIW during HS1 as suggested by Marchitto *et al.*, [2007] and others but the exceptionally aged signatures in the tropical oceans and the lack of an

aged signature along the flow path of AAIW out of the Southern Ocean necessitate direct release of a significant portion of the old carbon pool to intermediate depths in the tropics during HS1, fully depleting the Southern Ocean glacial carbon pool. Based on $\Delta^{14}\text{C}$ records from the deep south Pacific, which constrain the deep carbon reservoir between approximately 2000 and 3500 m water depth [Ronge *et al.*, 2016], we calculate that an increase in $\Delta^{14}\text{C}$ activity of 400‰ over this portion of the water column corresponds to the release of 339 GT of carbon from SPDW during HS1 [Skinner *et al.*, 2015, Supplementary Information Test S3]. Furthermore, in at least the vicinity of the Galapagos, ^{14}C depleted carbon likely reached the surface ocean and atmosphere during HS1; B-P ^{14}C offsets decrease by 2200 years during HS1, which supports enhanced mixing between surface and intermediate depth water masses relative to the LGM.

By the end of HS1 the Southern Ocean carbon reservoir was fully depleted (Fig. 5) and a second, distinct carbon source for the YD rise in atmospheric CO_2 is therefore required. Although no deep water radiocarbon signatures from regions outside the Southern Ocean are as exceptionally aged, PDW in the North Pacific was less well ventilated during the LGM than it is today; minimum $\Delta^{14}\text{C}$ values are observed between 2 and 3.6 km in the North Pacific reaching values as low as -350‰ [Cook and Keigwin, 2015]. Furthermore, and of key interest here, isolation of NPDW persists through HS1 [Cook and Keigwin, 2015; Jaccard and Galbraith, 2013; Galbraith *et al.*, 2007; Jaccard *et al.*, 2009]. Authigenic uranium concentrations in Gulf of Alaska sediments from 3.2 and 3.6 km reveal sustained suboxic conditions until at least 14.6 ka [Galbraith *et al.*, 2007; Jaccard *et al.*, 2009; Jaccard and Galbraith, 2013], and B-P ventilation ages are largest during the LGM and HS1 (Fig. 6).

Proxy evidence suggests that ventilation of NPDW was delayed 3.4 kyrs relative to SPDW [Galbraith *et al.*, 2007; Ronge *et al.*, 2016] (Fig. 6); however, given that NPDW is the source of SPDW today both PDW endmembers have undergone synchronous ventilation changes if the modern circulation pattern was maintained. To explain the delay in NPDW ventilation we therefore invoke an equatorward shift in the formation region of PDW. Today, “new” PDW forms north of 40°N; Circumpolar Deep Water flows north at abyssal depths, upwells to the base of the lower thermocline via vertical diffusion, and flows back to the Southern Ocean. During the last glacial period we suggest PDW formation occurred further south, thereby isolating a deep water mass in the far north Pacific until 14.6 ka when the formation region shifted poleward, ventilating the deep N. Pacific poleward of 40°N (Fig 7). Although this hypothesis is inline with a large body of proxy data [Galbraith *et al.*, 2007; Jaccard *et al.*, 2009; Jaccard and Galbraith, 2013; Cook and Keigwin, 2015; Ronge *et al.*, 2016], it still requires model simulations to test feasibility of the proposed circulation changes. Experiments that investigate the role of stratification and changing water mass geometries on PDW ventilation rates must therefore be the subject of future research. Galbraith *et al.*, [2007] link an abrupt change in North Pacific ventilation to a similarly abrupt resumption of the North Atlantic overturning circulation, although the dynamical link between the two basins remains unclear

Regardless of mechanism, the shift to enhanced deep North Pacific ventilation 14.6 ka occurs in phase with a sharp rise in atmospheric CO₂ concentrations by 10 p.p.m.v. [Grootes *et al.*, 1993; Marcott *et al.*, 2014]. Spikes in subarctic productivity and export production co-occur with the onset of enhanced deep water ventilation, which

indicates surface productivity was able to nearly keep pace with enhanced upwelling of nutrients and CO₂ [Galbraith *et al.*, 2007]. As a result, three-fourths of the upwelled ¹⁴C-depleted carbon was re-sequestered at intermediate depths [Galbraith *et al.*, 2007]. Intermediate depth radiocarbon records downstream, near Baja and within the EEP, reflect the influx of the re-sequestered but still ¹⁴C-depleted waters from the north; radiocarbon signatures trend towards more depleted values at the onset of the Bolling Allerod/ACR at both sites but was not able to reach the surface ocean or exchange with the atmosphere, at least in the EEP (Fig. 6) [Marchitto *et al.*, 2007; Lindsay *et al.*, 2015]. Galapagos B-P offsets indicate a return to near glacial levels of stratification and substantially reduced vertical mixing.

The 30 p.p.m.v. rise in atmospheric CO₂ during the YD interval reflects the subsequent release of re-sequestered North Pacific carbon under reduced stratification, and therefore, upwelling-favorable conditions. Between 17.2 and 14.1 ka the offset between NPDW $\Delta^{14}\text{C}$ and atmospheric $\Delta^{14}\text{C}$ dropped by 130‰, indicating the release of ~110 GT of carbon from NPDW at this time [Galbraith *et al.*, 2007; Skinner *et al.*, 2015, Supplementary Information Test S3]. Intermediate depth records from the Gulf of Alaska, close to the source, as well as the records from the tropical Pacific further afield (Baja and the Galapagos), find large increases in B-P ages coeval with the YD rise in CO₂ suggestive of an influx of older waters to the shallow subsurface 12.5 to 12.0 ka [Davies-Walczak *et al.*, 2014]. In the EEP, B-P offsets at the Galapagos site exhibit an abrupt shift from 2750 years to 150 in 1.5 kyrs, simultaneous to the rise in atmospheric CO₂ [Monnin *et al.*, 2004]. Although much smaller, B-P ¹⁴C offsets from 1000 m water depth along northern Peru Margin drop by 570 years. Stratification indices based on

$\delta^{18}\text{O}$, $\delta^{13}\text{C}$, and temperature gradients from the three sites also support a regional stratification minima during the YD [Bova *et al.*, 2015], while boron isotopes provide direct evidence for CO_2 outgassing (Fig. 8) [Martinez-Boti *et al.*, 2015]. We therefore suggest that during the YD and HS1 reduced stratification in the EEP provided a “carbon window” for old carbon to reach the surface ocean and atmosphere.

The radiocarbon records presented here cast doubt on the conventional view that AAIW was the primary conduit for the return of ^{14}C depleted carbon to the atmosphere during the last deglaciation [e.g. Marchitto *et al.*, 2007; Anderson *et al.*, 2009]. $\Delta^{14}\text{C}$ signatures at 600 m water depth on the Galapagos platform during the deglaciation are equivalent to those documented within PDW, the source of ^{14}C -deplete carbon, which suggests minimal dilution and short transport to the EEP. Preservation of such low $\Delta^{14}\text{C}$ signatures within AAIW along its circuitous route to the EEP, however, is not possible in model simulations with realistic circulations [Hain *et al.*, 2011]. In addition, intermediate depth radiocarbon records from along the flow path of AAIW to the EEP find no evidence for an aged water mass during either phase of atmospheric CO_2 rise during the last deglaciation [Pol-Holz *et al.*, 2010; Burke and Robinson, 2012]. If sourced from the Southern Ocean, those records closest to the AAIW formation region should document the most extreme ^{14}C depletions, much larger than what we observe further afield in the tropical Indo-Pacific, but the opposite is observed. We therefore suggest that direct injection of old carbon from its source, PDW, directly into the tropical Pacific subsurface is the best explanation for the intermediate depth ^{14}C anomalies detected along the Galapagos, Baja, and the Oman coasts during the two-part rise in atmospheric CO_2 during the last deglaciation (Fig. 7).

4. Methods

4.1 Area of study

Four sediment cores recovered from 3 intermediate water depths (373, 595-617, and 1023 m) in the EEP were recovered on the 2009 cruise of the R/V Knorr. These cores sample Equatorial Pacific Intermediate Water (EqPIW) a mixture of AAIW with contributions from PDW and SAMW [Fiedler and Talley, 2006; Bostock et al., 2010; 2013]. CDH 23 (03°44.95' S, 81°08.05' W, 373 m) and CDH 26 (03°59.16' S, 81°18.52' W, 1023 m), recovered along the Northern Peru Margin, record SAMW and EqPIW, respectively. At 600 m on the Galapagos Platform, GGC 43 (01°15.13'S, 89°41.07'W, 617 m) and CDH 41 (01°15.94'S, 89°41.88'W, 595 m) lie near the boundary between the two water masses.

SAMW and AAIW, the main components of EqPIW, are sourced primarily from the Southern Ocean [Bostock et al., 2010; 2013]. SAMW forms within the Subantarctic Zone during late winter convective overturning and travels a circuitous route from the Southern Ocean into the western equatorial Pacific before finally flowing eastward across the Pacific into the EEP [McCartney, 1975; Toggweiler et al., 1991; Rintoul and England, 2002; Qu et al., 2009; Bostock et al., 2010; 2013]. EqPIW is composed primarily of Antarctic Intermediate water (AAIW) with some contributions from PDW. AAIW forms closer to the continent than SAMW, south of the subantarctic front, in three primary formation regions: (1) southwest and east of New Zealand, (2) west of the East Pacific Rise, and (3) west of the Drake Passage [Bostock et al., 2013; Herraiz-Borreguero and Rintoul, 2011]. Intermediate waters in the EEP subsurface are sourced

primarily from the southeast Pacific site west of the Drake Passage, but its geochemical signature is distinct. EqPIW is significantly more saline, nutrient-rich, lower in oxygen and exhibits an older $\Delta^{14}\text{C}$ signature than Drake Passage AAIW, which requires mixing between AAIW and an older, more nutrient-rich waters mass, most likely PDW. North Pacific Intermediate Water is not an important source of intermediate water to the EEP today but some studies suggest a greater north Pacific influence in the past [e.g. *Max et al.*, 2014]

4.2 Radiocarbon Measurements

Paired planktonic and benthic radiocarbon measurements were made on four cores from the eastern equatorial Pacific. We use a splice between two cores from the Galapagos platform to extend the record from 600 m into the last glacial period [Bova et al., 2015]. In total, we made twenty-six paired measurements from the Galapagos cores, fifteen from CDH 23, and twelve from CDH 26. All radiocarbon dates were analyzed at the National Ocean Sciences Accelerator Mass Spectrometry facility at Woods Hole Oceanographic Institute. Planktonic dates were published previously by *Bova et al.*, [2015] for age control. In this study, 1.4 to 7.9 mg of benthic foraminifers were picked from the >150 μm size fraction (Appendix C). Particles and other debris were removed from each the sample with a fine hair brush before sonicating repeatedly in deionized water.

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Figures:

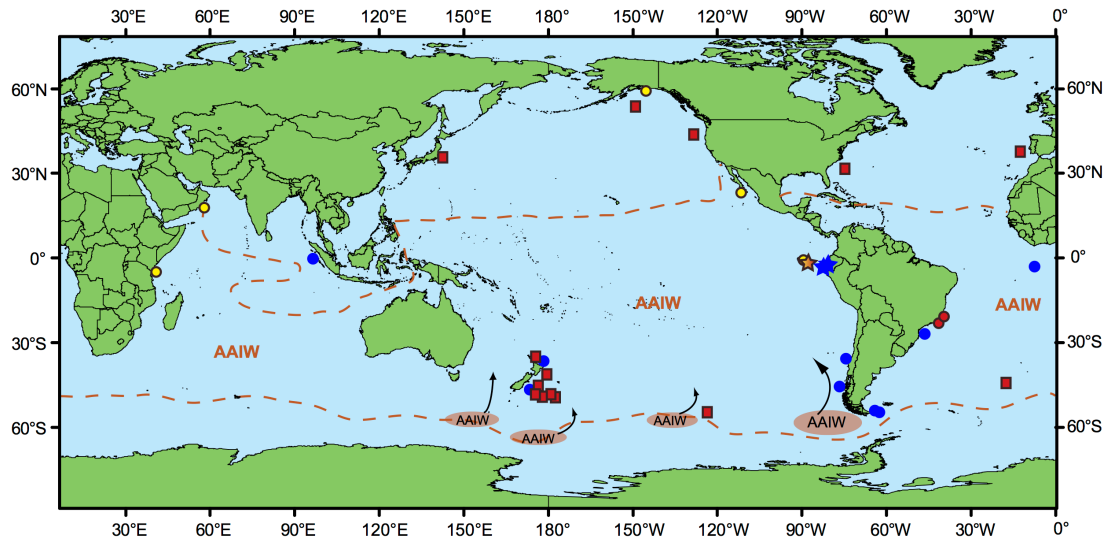


Figure 1. Overview of published intermediate and deep water radiocarbon records.

Locations of previously published benthic foraminiferal and coral radiocarbon records from intermediate water depths (circles) [Bryan *et al.*, 2010; Marchitto *et al.*, 2007; Lindsey *et al.*, 2015; Siani *et al.*, 2013; De Pol-Holz *et al.*, 2010; Davies-Walczak *et al.*, 2014; Sorter *et al.*, 2011; Romahn *et al.*, 2014; De Pol-Holz *et al.*, 2012; Burke and Robinson, 2012; Magini *et al.*, 2010; Rose and Sikes, 2010; Stott *et al.*, 2009; Kennett and Ingram, 1995], deep/abyssal water depths (squares) [Sikes *et al.*, 2016; Ronge *et al.*, 2016; Galbraith *et al.*, 2007; Lund *et al.*, 2011; Skinner *et al.*, 2010; Skinner *et al.*, 2014; Okazaki *et al.*, 2012; Robinson *et al.*, 2005; Keigwin, 2004] and three new records from intermediate depths in the EEP (stars, this study). Intermediate depth records that do (do not) exhibit excursions to depleted $\Delta^{14}\text{C}$ values during the deglaciation are plotted in yellow (blue). Deep and intermediate $\Delta^{14}\text{C}$ records that exhibit low $\Delta^{14}\text{C}$ values during

the last glacial period, indicative of the presence of a glacial carbon pool, are plotted in red. Orange dotted lines outline the extent of AAIW in each ocean basin [Talley et al., 2011]. Intermediate depth records that document an influx of old carbon during the last deglaciation are localized to the tropical Pacific, while those closer to the Pacific AAIW formation regions (pink ovals) do not. The record from 600 m on the Galapagos platform (this study) is the only record to exhibit evidence for old carbon during both the last glacial period and deglaciation.

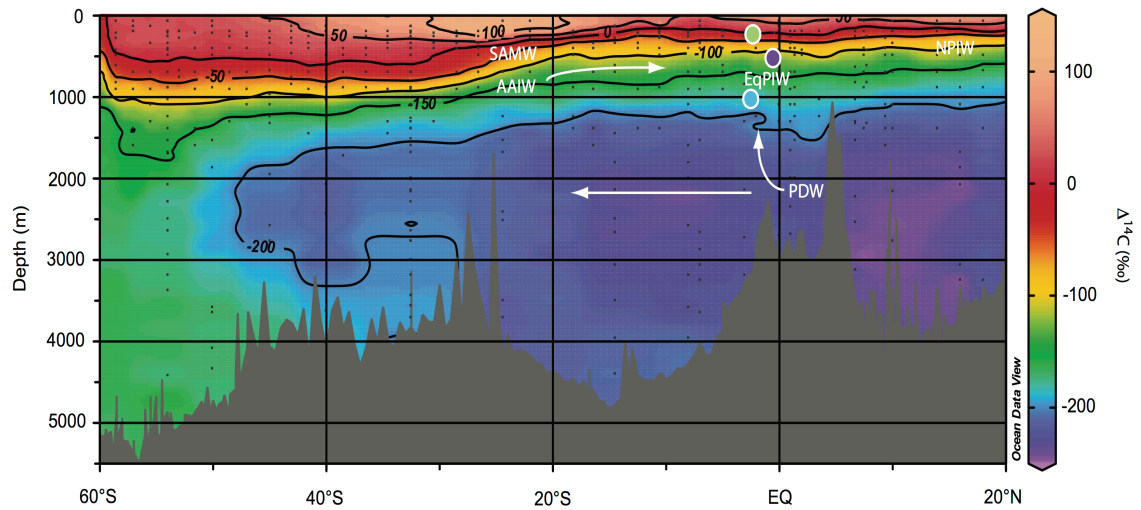


Figure 2. Modern Pacific $\Delta^{14}\text{C}$ concentrations along 90°W . [WOCE P19]. Locations of sediment cores CDH 23, CDH 26, and GGC43/CDH41 projected onto WOCE line P19. Core sites are bathed by a combination of SAMW and AAIW from the south with contributions from PDW. PDW is the oldest water mass in the global oceans today, with $\Delta^{14}\text{C}$ values as low as $\sim -250\text{‰}$ at its core. Figure generated using the ODV software.

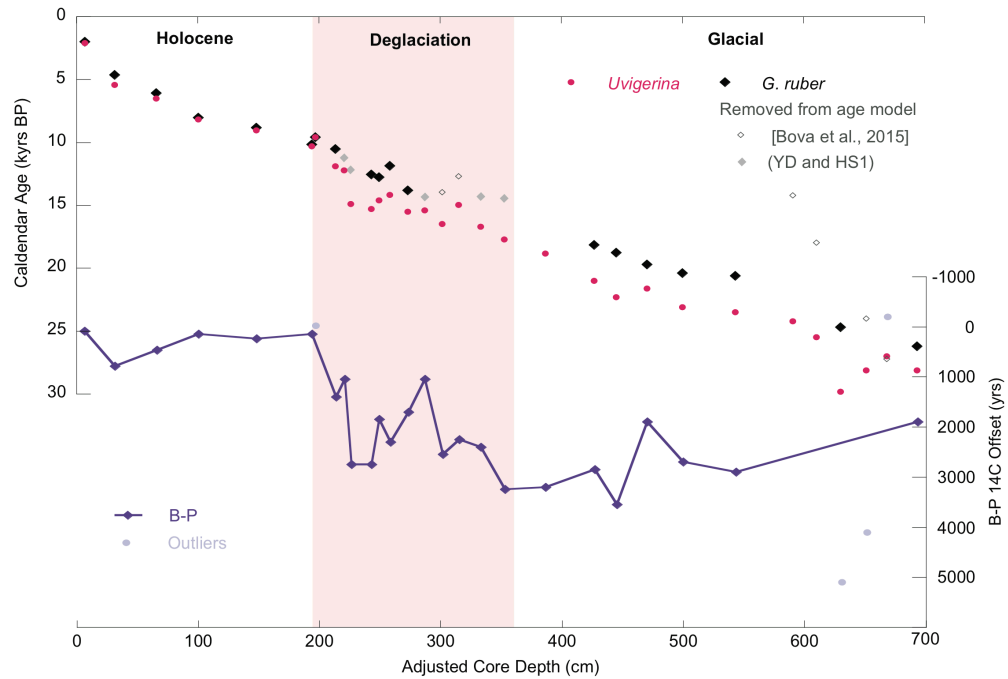


Figure 3. Paired planktonic and benthic foraminiferal ^{14}C dates. B-P offsets plotted versus adjusted core depth, a splice of GGC 43 and CDH 41. B-P offsets are largest during the last glacial period, which suggests strong stratification between surface and deep waters. B-P offsets decrease through the deglaciation towards a Holocene average of ~315 years. Seven planktonic ^{14}C dates are identified as outliers during the deglaciation likely due to reduced stratification conditions and increased upwelling of ^{14}C -depleted deep waters.

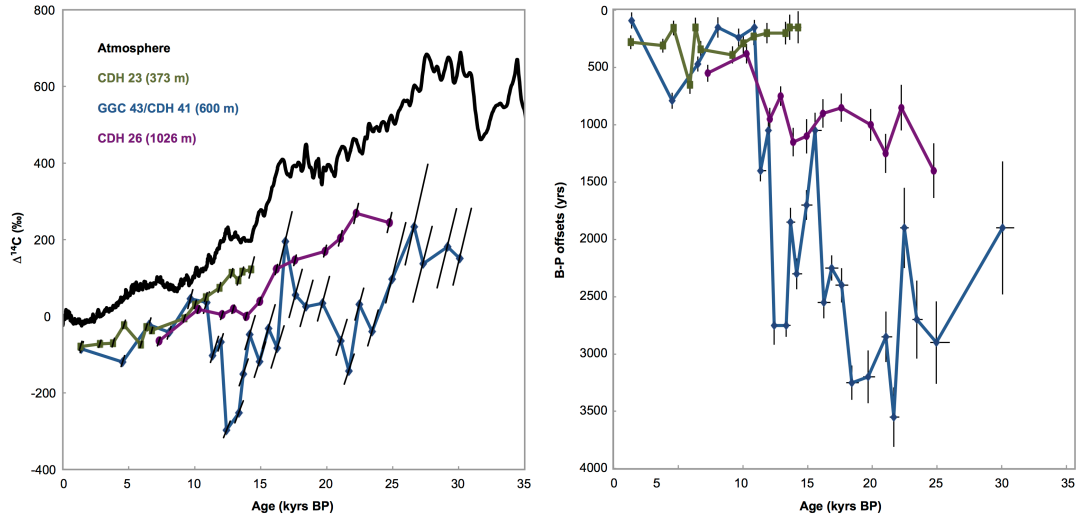


Figure 4. Radiocarbon data from 373, 600, and 1026 m water depth in the EEP.

(a.) $\Delta^{14}\text{C}$ concentrations at 600 m on the Galapagos Platform (purple) are exceptionally depleted during the last glacial period and deglaciation. The records from the N. Peru Margin (373 m = Green, 1026 m = Blue) maintain a near constant offset with atmospheric $\Delta^{14}\text{C}$ concentrations. Error bars reflect 2 sd in the age-depth model and are angled because the calculation for $\Delta^{14}\text{C}$ is contingent on calendar age [Blaauw, 2010].

(b.) B-P offsets exhibit similar trends to the $\Delta^{14}\text{C}$ signatures at the core sites (opposite in sign), with large offsets at 600 m during the LGM and deglaciation. At 1026 m water depth along the N. Peru Margin B-P ^{14}C age offsets are also larger during the LGM and deglaciation than during the early Holocene. Larger than modern B-P offsets imply reduced ventilation of intermediate water masses and/or enhanced stratification over the upper water column in the EEP relative to today. Error bars reflect 2 sd in the age-depth model in the x-direction and the sum of the measurement errors for both the planktonic and benthic ^{14}C dates in the y-direction.

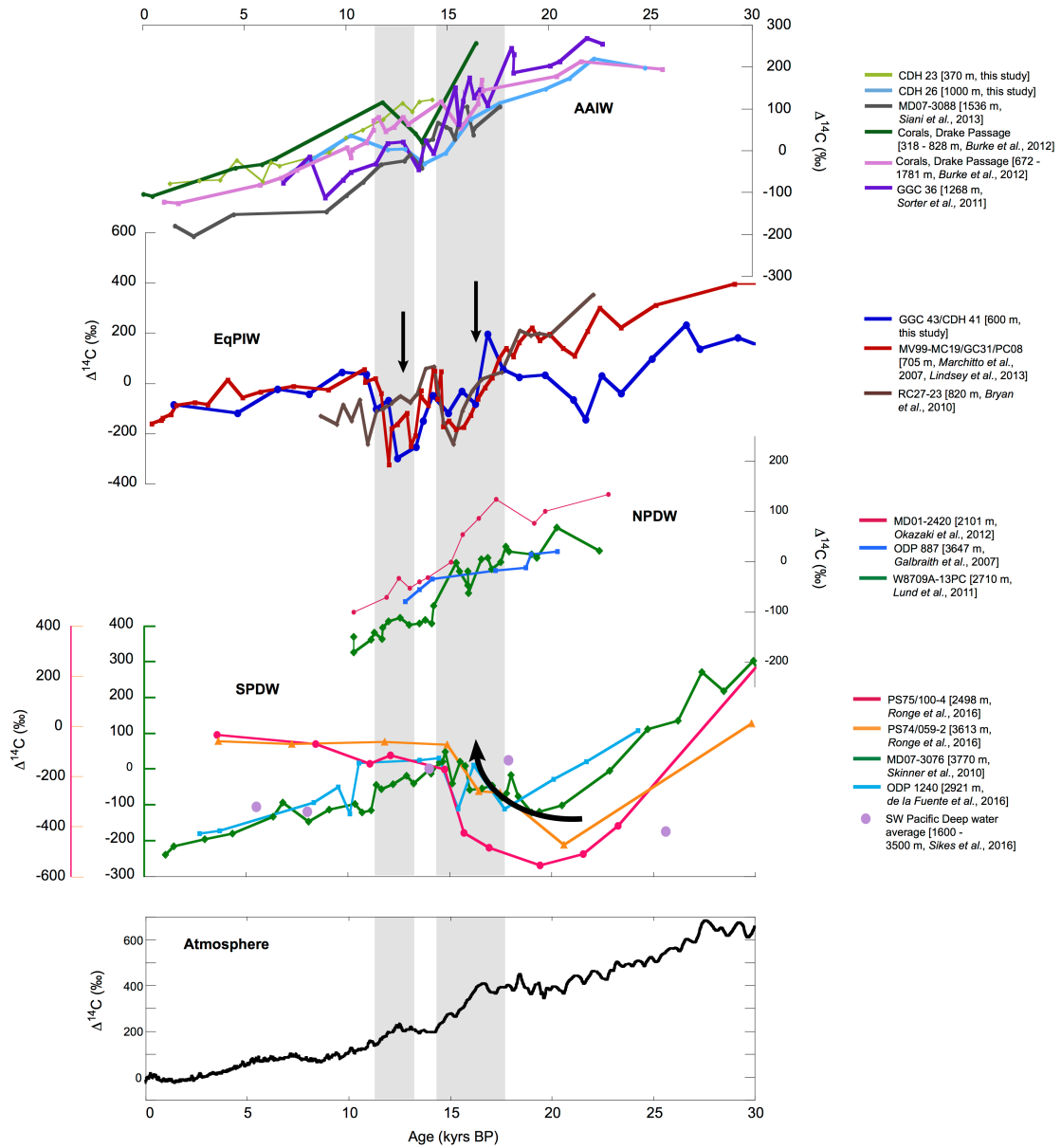


Figure 5. $\Delta^{14}\text{C}$ evolution over the past 30 kyrs by water mass. Radiocarbon records from AAIW depths in the Pacific Ocean maintain a relatively constant offset with the atmospheric $\Delta^{14}\text{C}$ concentrations over the past 30 kyrs [De Pol-Holz *et al.*, 2010; Burke and Robinson, 2012; Siani *et al.*, 2013]. The N. Peru Margin records from 373 and 1026 m water depth compare well to the AAIW $\Delta^{14}\text{C}$ history. Sites bathed by EqPIW exhibit negative $\Delta^{14}\text{C}$ anomalies during the deglaciation, similar to what we observed at 600 m

water depth on the Galapagos platform [Marchitto *et al.*, 2007; Lindsay *et al.*, 2015; Bryan *et al.*, 2010]. We exclude a record from marine sediment core VM21-30 recovered from 617 m, also on the Galapagos platform (Supplementary Text 2 and Supplementary Fig. 2) [Stott *et al.*, 2009]. $\Delta^{14}\text{C}$ values are exceptionally depleted in this record during the deglaciation, reaching nearly -600‰, and are not reconcilable with our record or any others from intermediate depths (Supplemental information Fig. 2). Finally, the radiocarbon history of PDW is characterized by low $\Delta^{14}\text{C}$ concentrations during the last glacial period, which recover to near modern values by the end of HS1 [Skinner *et al.*, 2010; Skinner *et al.*, 2015; Ronge *et al.*, 2016; Sikes *et al.*, 2016; Galbraith *et al.*, 2007; Lund *et al.*, 2011].

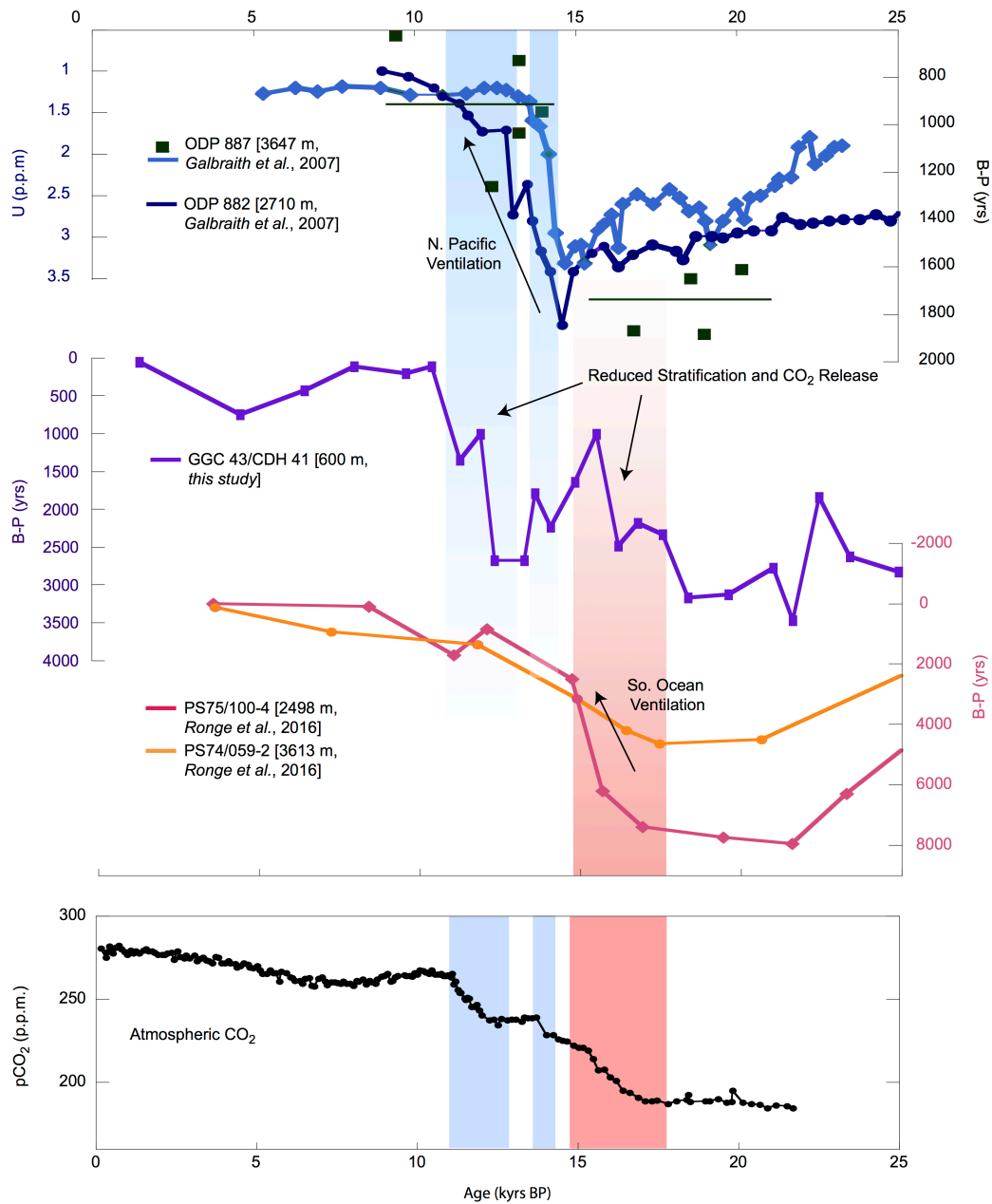


Figure 6. Release of the glacial carbon pool and the rise of atmospheric CO₂.

Radiocarbon and authigenic uranium records from the mid-depth Pacific ocean (2000-3500 m) indicate low levels of oxygen and radiocarbon in glacial PDW indicative of poor ventilation [Galbraith et al., 2007; Ronge et al., 2016]. ¹⁴C-depleted carbon stored in

South PDW was released during HS1, fueling the first rise of atmospheric CO₂ during HS1 [Monnin *et al.*, 2004; Ronge *et al.*, 2016]. In the EEP we observe a shift towards reduced B-P offsets, which suggests reduced stratification around the Galapagos Islands and enhanced vertical mixing of old carbon from South PDW to intermediate depths and the surface ocean. North PDW is ventilated ~14.6 ka in time with a 10 p.p.m.v. rise in atmospheric CO₂ levels, with the rest of the upwelled carbon re-squestered at depth by surface primary productivity and export to depth. Stratification breakdowns again in the EEP during the YD, which allows the North PDW carbon pool to reach the surface ocean atmosphere.

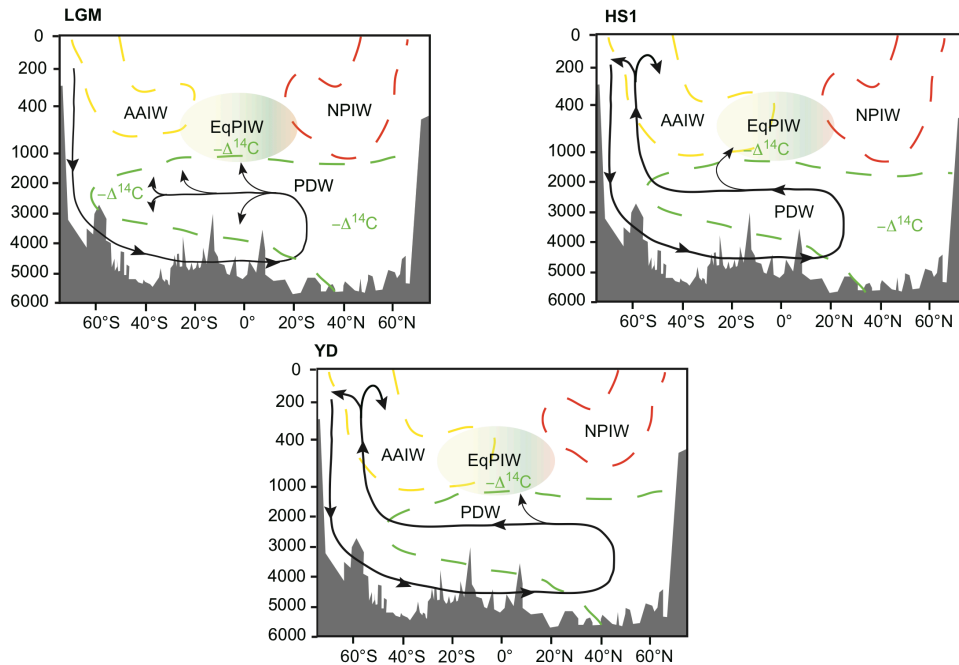


Figure 7. Schematic representation of the major phases of storage and release of the glacial carbon reservoir. (LGM, top left) Deep water circulation and upwelling in the Southern Ocean was reduced in the LGM Pacific, which allowed carbon to accumulate within PDW. Over time the ^{14}C -concentration of the carbon was reduced due to radioactive decay and isolation from the atmospheric source of ^{14}C . (HS1, top right) Enhanced upwelling around the Antarctic continent triggered the release of the PDW glacial carbon reservoir to the atmosphere. Old carbon was directly injected into the intermediate depth EEP and may have also upwelled to the surface ocean around the Antarctic continent. However, either only a small portion of the stored carbon upwelled in the Southern Ocean surface ocean or it quickly dissipated along the flow path of AAIW. (YD, bottom) During the YD the formation region of “new” PDW was shifted poleward to its present day position at 40°N , thereby ventilating ^{14}C -depleted carbon

stored in the far north deep Pacific during the last glacial period and HS1. Aged carbon was again injected into the EEP subsurface.

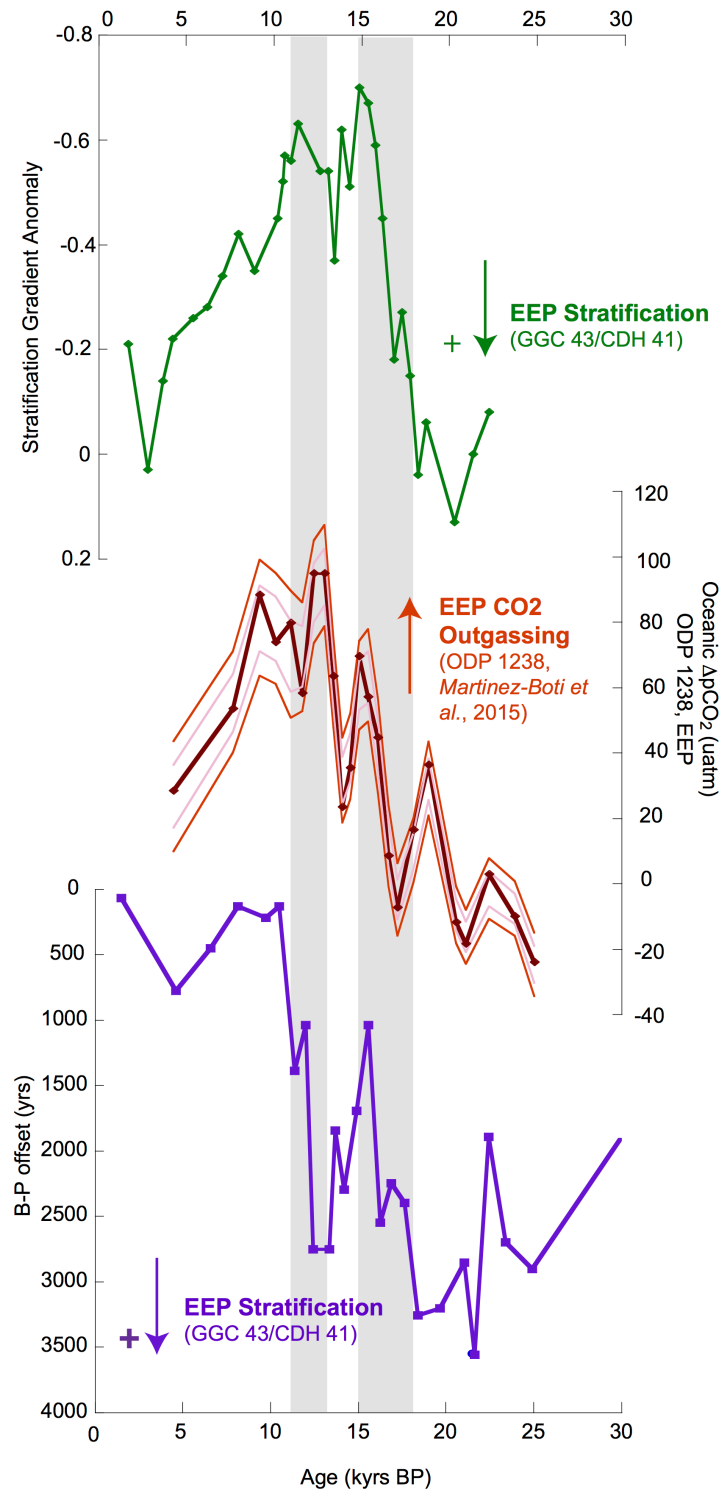


Figure 8. EEP stratification and CO₂ outgassing. ¹⁴C age offsets between 600 m depth on the Galapagos platform and the surface ocean (a.) compare well with other EEP

stratification records (c.) based on $\delta^{18}\text{O}$, $\delta^{13}\text{C}$, and temperature [Bova *et al.*, 2015]. Both stratification histories identify HS1 and the YD as periods of reduced stratification. (b.) Boron isotopes from nearby site ODP 1238 document CO_2 outgassing [Martinez-Botí *et al.*, 2015]. Periods of reduced stratification as documented here occur simultaneously with enhanced CO_2 outgassing at site ODP 1238, which suggests enhanced upwelling of PDW to the surface ocean during periods of reduced stratification in the EEP helped fuel the deglacial rise in atmospheric CO_2 .

Supplemental Information

Text S1. Reservoir Ages

Deep ocean $\Delta^{14}\text{C}$ calculations depend on knowing benthic foraminiferal ^{14}C ages and their corresponding calendar age. Implicit in this calculation is the assumption that the surface reservoir age does not change substantially over the past 25 kyrs. However, it is a practical certainty that surface reservoir ages have varied over the past 25 kyrs. There is strong evidence that upwelling rates and water column stratification have changed substantially over this time interval [e.g. *Koutavas et al.*, 2002; *Calvo et al.*, 2011; *Bova et al.*, 2015], potentially resulting in variable amounts of aged subsurface waters mixing with surface water and substantially altering surface reservoir ages [*de la Fuente et al.*, 2015]. Our data indicate that the largest reservoir ages (and thus the largest offset from the assumed constant reservoir age) occurred during the deglaciation. Enhanced upwelling of aged intermediate and deep waters are most likely to occur during intervals of reduced stratification. At the Galapagos site we observe reduced B-P offsets, indicative of reduced stratification, during the YD and HS1, coincident with the two-step rise in atmospheric CO_2 . Stratification indices based on $\delta^{18}\text{O}$, $\delta^{13}\text{C}$, and temperature records from the three sites confirm these intervals as periods of reduced stratification and enhanced vertical mixing (Fig. 7) [*Bova et al.*, 2015]. In contrast, *de la Fuente et al.* [2015] found that reservoir ages varied by 2000 years during the LGM at nearby site ODP 1240 ($0^\circ 01.31'\text{N}$, $86^\circ 27.76'\text{W}$). We find these reservoir age shifts counterintuitive; the largest surface reservoir ages, which the authors suggest reflect an enhanced influx of ^{14}C -depleted Pacific Deep Water to the surface ocean, occur during the LGM when B-P offsets are the largest. Large B-P offsets indicate strong stratification between surface and

deep water masses and reduced surface to deep mixing. Given the tentative and sparse nature of the tie points to the Greenland ice cores, the calculated reservoir ages are therefore suspect. We therefore suggest that, without independent age constraints, a constant 500 year reservoir age is the best approximation and is sufficient for interpretation and comparison to other subsurface ^{14}C records.

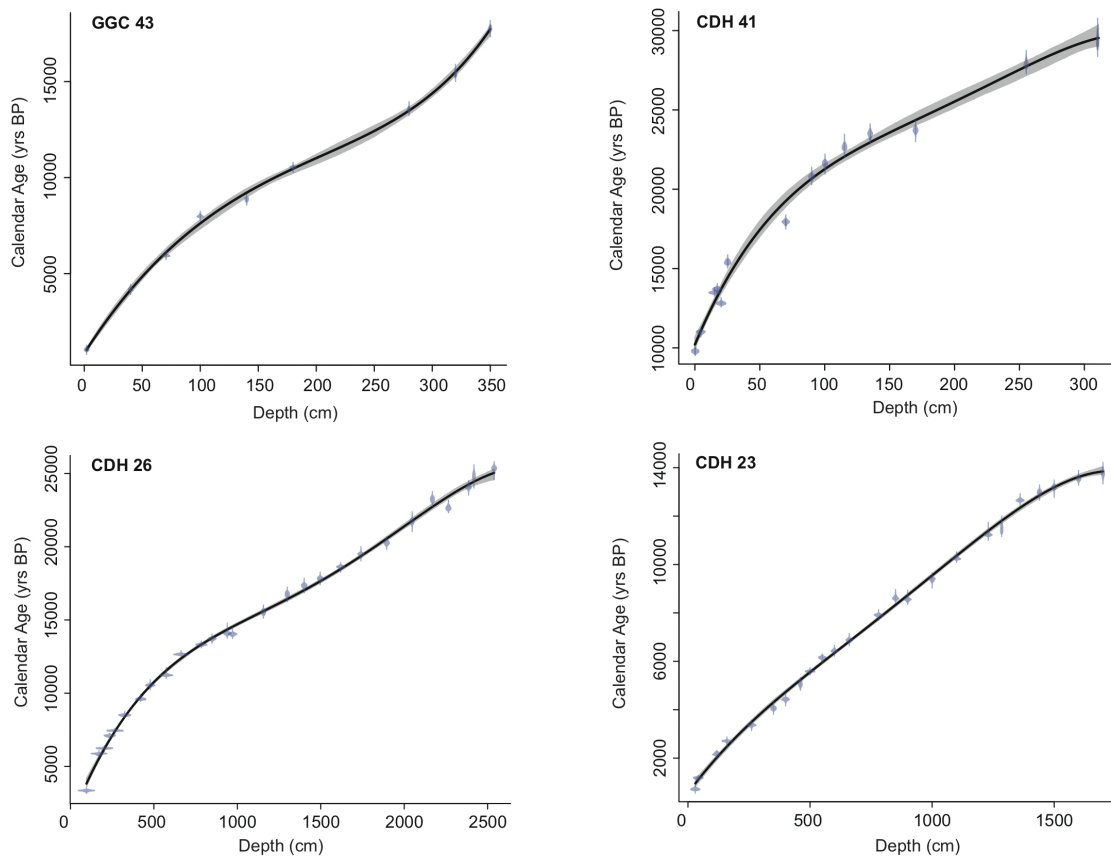
Text S2. Comparison to the $\Delta^{14}\text{C}$ record from VM21-30

Core VM21-30 was recovered just 5 km north of cores CDH 41 and GGC 43 but exhibits $\Delta^{14}\text{C}$ -concentrations as low as -600‰, approximately 300‰ lower than the minimum observed value in our record (Supplementary Fig. 2) [Stott *et al.*, 2009]. Stott and Timmermann [2011] suggest the anomalously old water masses detected at site VM21-30 can be explained by inputs of ^{14}C free geologic carbon via the destabilization of CO_2 -hydrates. Given the strong similarity between our ^{14}C record from intermediate depths on the Galapagos Platform with the record from off the coast of Baja, CA, however, it seems unlikely that CO_2 -hydrate destabilization would result in the same pattern at both locations simultaneously. We therefore do not consider the VM21-30 record in our analyses.

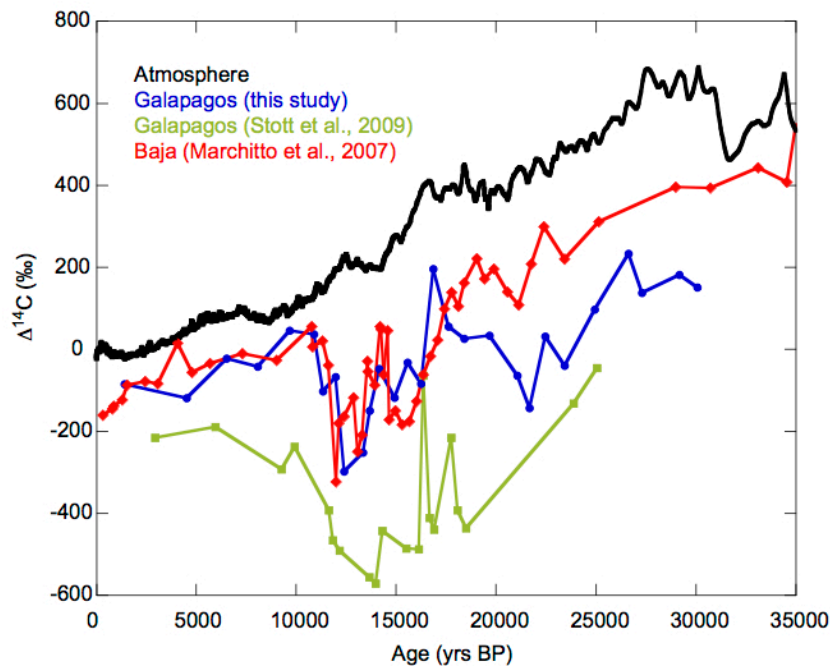
Text S3. Carbon storage in PDW

We base our estimates for the amount of carbon stored and released from within SPDW and NPDW on the empirical approach of Sarnthein *et al.* [2013] and the subsequent application by Skinner *et al.* [2015]. Skinner *et al.* [2015] demonstrate that seawater DIC increases by $\sim 1.22 \mu\text{mol/kg}$ per ‰ decrease in radiocarbon concentration. We

approximate the volume of water in the Pacific ocean between 2000 and 3500 m as $1.13 \times 10^{17} \text{ m}^3$ and halve this number to approximate the Northern and Southern Pacific Deep Water volumes. During HS1, SPDW and NPDW $\Delta^{14}\text{C}$ concentrations rose by approximately 400‰ [Ronge *et al.*, 2016] and 130‰, respectively. An increase of 400‰ corresponds to an increase in deep water DIC by 488 $\mu\text{mol/kg}$ and thus the release of 339 GT of carbon from SPDW. A 130‰ increase, as observed within NPDW during the YD, corresponds to a roughly 158.6 $\mu\text{mol/kg}$ increase in DIC and the release of 110 GT of carbon. Approximately two-thirds of the deglacial rise in atmospheric CO_2 occurred during HS1 and one-third during the YD. Our calculations demonstrate that roughly two-thirds of the carbon released from the deep Pacific came from SPDW (339 GT carbon) during HS1 and one-third from NPDW (110 GT carbon) during the YD and therefore correspond well to the observed changes in atmospheric CO_2 concentrations.



Supplementary Figure 1. Core chronologies. ^{14}C dates using the *Clam* version 2.2 age-modeling software package for R [Blaauw, 2010]. All ^{14}C dates were converted to calendar ages using the Marine13 radiocarbon age calibration curve [Reimer *et al.*, 2013]. The shaded region represents the 95% confidence window. Because B-P offsets at the Galapagos site vary wildly over the last 35 kyrs, implying significant change in vertical stratification near the core site and variable rates of deep/intermediate water upwelling to the surface ocean, we remove five ^{14}C dates from the original core chronologies of Bova *et al.* [2015] during intervals of decreasing B-P offsets to avoid possible changes in the surface reservoir age. Together with dates removed in the original chronology a total of ten ^{14}C dates from the Galapagos cores were removed [Appendix C and Fig. 3 in the main text].



Supplementary Figure 2. Comparison to previously published benthic $\Delta^{14}\text{C}$ record

from the Galapagos Platform. A paired benthic and planktonic ^{14}C record from core VM21-30 recovered from 617 m water depth on the the Galpagos platform (green symbols) [Stott *et al.*, 2009] exhibits a radically different benthic foramiferal $\Delta^{14}\text{C}$ history than that observed at site GGC 43/CDH 41 (this study, blue symbols) and off the Baja Peninsula [Marchitto *et al.*, 2007]. Core VM21-30 was recovered just 5 km north of cores CDH 41 and GGC 43 but exhibits $\Delta^{14}\text{C}$ -concentrations as low as almost -600‰, approximately 300‰ lower than the minimum observed value in our record.

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CHAPTER FOUR

Relationships between thermocline depth, sea surface temperature, and productivity in the Eastern Equatorial Pacific: no role for ENSO since the last glacial period

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Abstract

The Eastern Equatorial Pacific (EEP) is a complex environment, with important feedbacks to the global carbon cycle and ocean heat content. The shallow thermocline, characteristic of the EEP, connects the EEP subsurface with the surface ocean and atmosphere, stimulating high rates of primary production as well as rapid heat and gas exchange with the atmosphere. Today, the El Niño Southern Oscillation (ENSO), an ocean-atmosphere interaction that originates over a small latitudinal band of the tropical Pacific, exerts primary control on the depth of the EEP thermocline and is the second largest source of variability in the region on human timescales, after the seasonal cycle. Predicting the response of the EEP thermocline and ENSO to alternate climate background states is therefore of primary concern. In this study, we assess the response of the EEP to climate change since the last glacial period. We reconstruct a 30 kyr history of SST, thermocline depth, and phytoplankton productivity at four sites in the EEP. We find that the modern relationships between sea surface parameters, characteristic of ENSO, are not maintained over the past 30 kyrs. Mixed layer properties were likely affected primarily by the local wind-driven forcing, while remote forcing regulated thermocline depth. Thus, though ENSO may have been active over the past 30 kyrs, its influence on EEP thermocline depth was overshadowed by the influx of subsurface temperature anomalies from the extratropics.

1. Introduction

El Niño, and its counterpart, La Niña, are the most important source of interannual climate variability on the planet. The phenomena, termed the El Niño Southern Oscillation (ENSO), lead to large changes in the heat distribution across

equatorial Pacific with important consequences for global atmospheric circulation patterns [e.g. *Philander*, 1985]. Predicting the response of ENSO to future global warming is therefore of primary concern to policy makers. Thus far, simulations with coupled general circulation models project conflicting ENSO responses to warm scenarios: some model results indicate change in the zonal mean thermocline depth of the equatorial Pacific [*DiNezio et al.*, 2009] while others predict sustained “ENSO-like” variation in its tilt [*Vecchi and Soden*, 2007]. Given the poor constraints available, records from the past are necessary to test model estimates of the EEP thermocline response under alternate climate background states.

The ENSO cycle is characterized by specific relationships between thermocline depth, sea surface temperature, and primary productivity, each of which can be independently quantified in paleoceanographic records [e.g. *Cane et al.*, 1997; *Pennington et al.*, 2006]. A shallow thermocline allows the surface wind to more efficiently entrain and upwell cold, nutrient-rich sub-thermocline water masses. Consequently, during El Niño when the thermocline is deep, warm SSTs are coupled with reduced primary production, and during La Niña when it is shallow, SSTs cool and primary production increases [e.g. *Philander et al.*, 1995; *Barber et al.*, 1996]. This is the classic El Niño Southern Oscillation (ENSO) paradigm- one paleoceanographers extend to timescales of 10^2 - 10^6 years [*Beaufort et al.*, 2001; *Koutavas et al.*, 2002; *Stott et al.*, 2002; *Wara et al.*, 2005; *Pena et al.*, 2008]. A recent synthesis of Holocene SSTs, using both Mg/Ca and alkenone paleo-SST proxies, finds that equatorial Pacific SSTs were in a La Niña-like distribution and therefore infer reduced ENSO variability prior to 2 ka [*Gill et al.*, 2016]. But do SST patterns adequately capture dynamics in the eastern equatorial

Pacific that play out on timescales of decades to millennia? That is, can ENSO serve as a model for shifts in the EEP mean state under differing boundary conditions? A number of theoretical paleoceanographic observations [*Seager and Murtugedde*, 1997; *DiNezio et al.*, 2011] suggest that the answer may be “no”.

On centennial and longer scales, a primary forcing on thermocline depth, SST, and paleoproductivity may come from the extratropics. Subsurface variability reflecting anomalies sourced from the mid- [*Liu and Philander*, 1995; *Gu and Philander*, 1997; *Lu et al.*, 1998; *Boccaletti et al.*, 2004] and high-latitudes [*Toggweiler et al.*, 1991; *Andreasen et al.*, 2001] might play an essential role in setting the background stratification of the EEP on centennial and longer scales. These signals emerge as rapid changes in denitrification [*Chazen et al.*, 2009] and subsurface water density along the Peru margin [*Bova et al.*, 2015]. In the case of positive nitrogen isotope anomalies (increased strength of the oxygen minimum zone (OMZ)), these coincide with *minima* in inferred paleoproductivity, indicating that increased residence time of “shadow-zone” waters, not a wind-driven upwelling increase in carbon flux and oxygen demand, trigger centennial pulses in the OMZ. The observed subsurface anomalies occur too deep to be forced from the surface by mechanisms associated with ENSO.

In this study, we apply independent proxies to determine the thermocline depth, SST, and productivity at four EEP locations over the past 30 kyrs to provide an essential test of the hypothesis that millennial to orbital scale changes in the oceanography of the region can be regarded as “scaled-up” El Niño or La Niña conditions. The transition from the most recent glacial period into the current interglacial provides an excellent opportunity to test the EEP upper ocean response to glacial boundary conditions as well

as a rapidly warming climate. We provide SST and productivity estimates along with the first quantitative regional reconstruction of the depth of the thermocline based on the relative abundance of the coccolithophore *Florispheara profunda*. Based on the observed correlations between SST, productivity, and thermocline depth we reject ENSO as the dominant source of variability in the EEP since the last glacial period. In addition, we find that SST is not a straightforward indicator of thermocline depth and is therefore a necessary, but not sufficient, parameter for determining whether on millennial and longer timescales the EEP can shift between states resembling dominantly El Niño, La Niña, or a state similar to neither.

2. Oceanographic Setting

2.1 Mean Hydrography

Circulation patterns in the EEP upper water column are highly complex and result in sharp chemical and physical gradients. The region's most prominent feature, the Equatorial Cold Tongue (ECT), centered at 1°S, is composed of three intersecting but dynamically distinct upwelling systems: the Peru Chile Upwelling (PCU), the Equatorial Upwelling (EU), and the Galapagos Upwelling (GU) systems [e.g. *Fiedler and Talley*, 2006]. Along-shore winds drive surface waters away from the South American coast in the PCU system, bringing cool, nutrient-rich subsurface waters from depth to the surface ocean [*Pennington et al.*, 2006]. The EU system is sustained by easterly winds that drive waters away from the equator as the Coriolis force changes sign across the equator. Westward advection of upwelled waters from the Peru Current connect the two upwelling systems creating a tongue of cool water just south of the equator in the EEP, known as Equatorial Surface Water (ESW). Finally, collision of eastward flowing subsurface

Equatorial Undercurrent (EUC) with the Galapagos Islands drives perennial upwelling west of the islands [Kessler, 2006]. Surface water properties in the ECT are thus determined by advection of cool waters from the Peru Current and the strength of upwelling within each system [Wyrtki, 1966; 1981; Fiedler and Talley, 2006].

A sharp thermal and salinity gradient along 2°N separates the ECT from the eastern Pacific warm pool (EPWP) [Fiedler and Talley, 2006]. The EPWP is the second largest tropical warm pool on the planet and arises through a combination of weak wind-driven mixing and a seasonally large net heat flux [Wang and Enfield, 2001]. The sharp SST gradient between the ECT and the warm pool is maintained by the influx of cool waters primarily via local upwelling along the equator with smaller contributions from zonal advection of waters from the PCU system to the east [Kessler *et al.*, 1998; Kessler, 2006].

The eastward flowing EUC, centered at 80 m depth in the EEP, supplies the PCU, the GU, and the EU systems. It is composed primarily of modified subtropical mode water with contributions from southern ocean mode and intermediate waters (SOIW) sourced from the southeast Pacific and northeast of New Zealand [Toggweiler *et al.*, 1991; Qu *et al.*, 2009; Bostock *et al.*, 2010] and small additions from North Pacific Intermediate Water (NPIW). Tracer experiments [Qu *et al.*, 2010] demonstrate that intermediate waters do eventually upwell to the surface ocean in the EEP but have minimal impacts on regional SSTs through at least the Holocene [Kalansky *et al.*, 2015]. Nevertheless, SOIW and NPIW are the primary macronutrient sources to the EEP; SOIW and NPIW are both nitrate- and phosphate-rich but silicate is sourced almost exclusively from the north [Sarmiento *et al.*, 2004]. Competition between northern and southern

intermediate waters can be diagnosed as a dichotomy of silica-rich thermocline waters from the north Pacific versus silica poor waters from the south Pacific [*Dubois et al.*, 2010; *Kienast et al.*, 2006; *Sarmiento et al.*, 2004] and north-south shifts in the relative abundance of siliceous organisms in the EEP.

2.2. Seasonal and Interannual Variability

Seasonal migration of the Intertropical Convergence Zone (ITCZ) is the primary driver of variability in the ECT in the historical record [e.g. *Fiedler and Talley*, 2006]. The seasonal amplitude of SST variability in the ECT is $\pm 1-3^{\circ}\text{C}$, with the coldest temperatures observed in September and October, when the Intertropical Convergence Zone reaches its most northerly position and strong, easterly winds over the region drive a mean shoaling of the thermocline south of the ITCZ by about 20 m [*Fiedler and Talley*, 2006]. A shallow thermocline acts as positive feedback to surface cooling, allowing winds to more efficiently mix cool, nutrient-rich subsurface waters to the surface ocean [*Kessler*, 2006]. Because the delivery of macronutrients exerts primary control on rates of production in the EEP today, phytoplankton are expected to respond in phase with the seasonal cycle. However, phytoplankton in the EEP do *not* exhibit strong seasonality and, in fact, in the northern PCU system phytoplankton exhibit some of the highest rates of production during boreal winter when upwelling rates are at a minimum [*Chavez*, 1995; *Pennington et al.*, 2006]. Light limitations resulting from stratus cloud formation and fog over cool surface waters or dilution by advection and mixing may explain this surprising relationship.

ENSO is the second leading cause of variation in upwelling strength, thermocline depth, and SST in the EEP over the time period of historical observations, and the

primary source of variability in phytoplankton growth [Pennington *et al.*, 2006; Wang and Fiedler, 2006]. The influence of ENSO is strongest in the EU and coastal boundary regions ($\pm 1-2^{\circ}\text{C}$) and much less so in the warm EPWP [Fiedler and Talley, 2006]. Phytoplankton production decreases substantially during El Niño events across the entire EEP and in coastal regions ecological shifts from dominantly diatoms and larger phytoplankton communities to smaller phytoplankton, typically observed in the open-ocean, have been observed during particularly strong events [e.g. Chavez, 1989; Chavez *et al.*, 1991].

3. Site Locations

Processes that might affect the evolution of the EEP over the past 30 kyrs include local wind-driven mechanisms, such as ENSO, and remote forcing from the extratropics. The spatial response of the EEP to each mechanism is different and can help distinguish the relative role of local versus remote controls on EEP oceanography. We therefore reconstruct thermocline depth, SST, and primary production at multiple sites spread across the region. Sediment cores used in this study were recovered from four distinct hydrographic regimes in the EEP: (1) the EPWP (ME0005A-15MC/17JC), (2) the open-ocean ECT (VNTR01-13GC), (3) the PCU system (CDH 23/CDH 26), and (4) the GU system (GGC 43/CDH 41) (Figure 1, Table 1).

4. Methods

4.1 Age models

Core chronologies were constructed by applying polynomial fits to AMS ^{14}C dates of planktonic foraminifera using the Clam age model software of Blaauw [2010] (Figure 2). Radiocarbon measurements were made at the National Ocean Sciences

Accelerator Mass Spectrometry facility at Woods Hole Oceanographic Institute.

Planktonic foraminifera were picked from the >150 μm size fraction. All ^{14}C dates were converted to calendar ages using the marine13 radiocarbon age calibration curve [Reimer et al., 2013]. We assume constant reservoir ages at each of the sites based on the nearest modern measurements from the Marine Reservoir Age database (Table 1) [Taylor and Berger, 1967; Druffel, 1981; 1995; 2004; Jones et al., 2007; Etayo-Cadavid et al., 2013].

Age models for sites CDH 23 (373 m) and CDH 26 (1026 m) are based on 24 and 25 AMS ^{14}C measurements, respectively, made on the thermocline dwelling foraminifer, *Neoglobobulimina dutertrei*. Cores CDH 23 and CDH 26 are spliced together at 13.7 ka to provide a complete record spanning the last 30 ka. Planktonic ^{14}C dates at the Galapagos sites were measured on *Globobulimina Ruber*. We modify the published chronologies of Bova et al., [2015] for the Peru Margin and Galapagos cores by using the marine13 calibration curve and the Clam Software package to characterize age model uncertainty instead of the Fairbanks calibration program. In addition, we remove seven planktonic ^{14}C dates as outliers from the chronology for cores GGC 43 and CDH 41 (617 and 595 m, respectively) [see Chapter 3, Results].

Age models for ME0005-15MC/17JC and VNTR01-13GC are based on 5 (*G. Ruber*) and 9 (*N. dutertrei*) AMS ^{14}C measurements, respectively. In addition, oxygen and carbon isotopic signatures of benthic foraminifera from each site were used to confirm the initial ^{14}C stratigraphy by comparison to LR04 benthic stack [Lisiecki and Raymo, 2005]. Isotopes were measured on *Uvulimina peregrina* and *Cibicides wuellerstorfi* at site VNTR01-13GC. *C. wuellerstorfi* was not abundant at site ME0005A-17JC so only *U. peregrina* species were measured. Based on the radiocarbon and oxygen

isotope data, we identify two identical Holocene sections in core VNTR01-13GC that must represent either slumping or duplication in coring. Because the top Holocene section appears more complete, we splice the top 52 cm of the core, spanning 1.35 to 16.6 ka, to early deglacial sediment beginning at 90 cm, and remove measurements from the middle 36 cm of the core from the dataset.

4.2 Biomarkers

Organic biomarker records were analyzed in cores GGC 43/CDH 41, VNTR01-13GC, and ME0005A-15MC/17JC on an Agilent 6890 Gas Chromatograph with a Flame Ionization Detector [GC-FID]. Samples from the Galapagos cores, GGC 43 and CDH 41, were analyzed only for alkenone biomarkers. Alkenone records from the N. Peru Margin sites (CDH 23 and CDH 26) were published previously by *Bova et al.*, [2015]. We use the calibration equation of *Müller et al.*, [1998] to calculate paleo-SSTs from alkenone distributions.

Lab analytical error is equivalent to $\pm 0.1^{\circ}\text{C}$ based on 55 replicate analyses of a laboratory standard. Approximately 10% of our analyzed samples were run in duplicate with an average $\text{U}^{\text{K}'}_{37}$ -based temperature analytical reproducibility of $\pm 0.05^{\circ}\text{C}$. Brassicasterol abundances were also characterized by GC-FID analysis. Thirty-nine brassicasterol samples were run in duplicate with an average reproducibility of ± 6.5 ng/g. For a more detailed description of our methods please refer to *Chapter 5*.

4.3. Nannofossil Abundances

Twenty-seven core-top samples located between 15°N and 15°S and east of 110°W were analyzed for coccolith abundances (Figure 5). Smear slides were prepared using Norland Optical Adhesive 81. Coccolith identification was carried out using a

Leica DMLSP polarized microscope at 1250x magnification. All slides were analyzed using simple relative abundance counting. A minimum of 300 species were identified per slide, which assures identification of any taxon whose abundance is greater than 1% of the total population [Thierstein *et al.*, 1977].

Molfino and McIntyre [1990] and Bolton *et al.* [2010] previously identified a negative relationship between the abundance of *F. Profunda* and thermocline depth but a quantitative relationship has not yet been established for the EEP. We calculate the relative abundance of *F. profunda* according to the following formula adapted from Flores *et al.*, [2000]. $N\ ratio = \%G. oceanica + \%E. huxleyi / (\%G. oceanica + \%E. huxleyi + \%F. profunda)$. This index computes the abundance of *F. profunda* relative to the most abundant surface dwelling coccolithophores, *G. oceanica* and *E. huxleyi*.

The *N ratio* is highly sensitive to dissolution biases because of the differential dissolution behavior of *Emiliania huxleyi* and *Gephyrocapsa oceanica* relative to *F. profunda*. *E. huxleyi* and *G. oceanica*, are delicate and therefore more prone to dissolution than *F. profunda*. To avoid dissolution biases we therefore only use core-tops that lie above the modern lysocline (3700 to 4000 m) [Weber *et al.*, 1995]. Even above the nominal lysocline, however, dissolution can be still a problem. We therefore assess coccolith preservation using the CEX' index following Boeckel and Baumann [2004]: $CEX' = (\% E. huxleyi + \% G. oceanica) / (\% E. huxleyi + \% G. oceanica + \% C. leptoporus)$. The index compares the preservation state of *Emiliania huxleyi* and *Gephyrocapsa oceanica*, relative to the robust species, *Calcidiscus leptoporus*. Any sample with a CEX' index less than 0.63 is removed from the *N ratio* calibration.

Coccolithophorid relative abundances were calculated down-core at three EEP locations: ME0005A-15MC/17JC, GGC 43/CDH 41, and VNTR01-13GC. Slides were prepared and analyzed in the same manner as the core-top samples. Counts were completed approximately every 2 cm over the top 60 cm in core ME0005A-15MC/17JC, every 10 cm over the top 280 cm in core GGC 43/CDH 41 and every 6 cm over the top 200 cm in core VNTR01-13GC to provide a measurement approximately every 1 to 1.5 kyrs over the past 30 kyrs. The average standard deviation in the *N ratio* based on five duplicate counts is 0.033. Alkenone and brassicasterol biomarker concentrations were analyzed for all counted intervals.

5. Results

5.1 Alkenone Sea Surface Temperature Reconstructions

The SST history of the EEP across the past 25 kyrs, as reconstructed by alkenones is geographically heterogeneous. The timing of minimum SSTs at the end of the last glacial period and the amplitude of deglacial warming differs by location. At the northernmost site (ME0005A-15MC/17JC, 4.61°N) SSTs decrease during the last glacial period, reach a minimum at 20.0 ± 0.4 ka, and then increase through the deglaciation and into the early Holocene by 1.15 ± 0.05 °C. Early Holocene SSTs at this site are comparable to those observed during the last glacial period between 33 and 35 ka. At the Galapagos (1.25°S) we observe a broadly similar pattern, with minimum SSTs occurring slightly later at 18.0 ± 0.3 ka and the amplitude of LGM to Holocene warming slightly larger, at 2.04 ± 0.05 °C. The higher resolution of the record enables observation of a two-part warming during the last deglaciation during both periods of SH warming (11.2

to 12.3 & 16.4 to 18 ka). Holocene SSTs stabilize during the late Holocene at 24.01 ± 0.05 °C.

The two southernmost sites exhibit markedly different SST histories both from each other and from the Galapagos and northernmost site. At 3.09°S (VNTR01-13GC) we observe a minimum in SSTs (20.88 ± 0.05 °C) during the last glacial period at 25.8 ± 0.4 ka. SSTs rise by 2.75 ± 0.05 °C towards the onset of deglaciation at 19.8 ± 0.3 ka and exhibit high variability, with multiple 1 to 1.5 ± 0.05 °C temperatures swings in less than 1 kyrs. Temperatures stabilized during HS1 at 22.70 ± 0.05 °C before rising another 2.23 ± 0.05 °C between 15.5 ± 0.4 and 13.2 ± 0.4 ka. Unlike the Galapagos record, SSTs decrease at this location during the early Holocene. SSTs reach a minimum temperature of 23.48 °C ± 0.05 °C by the late Holocene. The record from our southernmost site along the Peru Margin (3.99°S) exhibits the latest onset for warming during the deglaciation, 14.9 ± 0.1 ka. SSTs rise 4.15 °C from 19.81 ± 0.05 to 23.96 ± 0.05 °C beginning at 14.9 ± 0.1 ka. Holocene SSTs here decrease towards present day but exhibit millennial-scale variability.

5.2 Productivity Estimates

We analyzed C₃₇-alkenones and brassicasterol concentrations to characterize coccolithophorid and diatom production at four and two sites, respectively, in the EEP: ME0005A-15MC/17JC (alkenones and brassicasterol), CDH 41/GGC 43 (alkenones only), VNTR01-13GC (alkenones and brassicasterol), and CDH 23/CDH 26 (alkenones only). Alkenone concentrations are a proxy for haptophyte or coccolithophorid production and brassicasterol concentrations have been shown to trace diatom production rates [e.g. *Higginson et al.*, 2004; *Calvo et al.*, 2011].

Biomarker records from across the EEP exhibit strong spatial heterogeneity. Open ocean sites (ME0005A-15MC/17JC and VNTR01-13GC) are most similar with maxima in both C₃₇-alkenone and brassicasterol abundances at the end of the last glacial period between approximately 15 and 31 ka. C₃₇-alkenone and brassicasterol abundances at sites ME0005A-15MC/17JC and VNTR01-13GC are positively correlated, with r-squared values of 0.83 (p<0.01) and 0.38 (p<0.01), respectively [Chapter 5, Figure 6].

Productivity records from the Galapagos and Peru Margin sites are unlike those from the open ocean regions of the EEP. High alkenone abundances are observed along the Peru Margin during the last glacial period, similar to the open ocean sites, but are also high during the Holocene. C₃₇-alkenone abundances increase at the onset of the Holocene until 6.4 ka and then decrease towards present day. In addition, alkenone abundances along the Peru Margin exhibit increased variance in the mid- to late-Holocene. At the Galapagos, alkenone abundances decrease between 27.9 and 22.0 ka from 6300 to 3200 ng/g and remain stable at 3900 ng/g until 11 ka. C₃₇-alkenone abundances increase from 11 ka to present day.

5.3 Quantitative Thermocline Depth Calibration

The *N ratio* is negatively correlated ($R^2=0.92$, $p<0.001$, Figure 5) with modern thermocline depth [Locarini *et al.*, 2013], defined here as the midpoint of the depth interval with the maximum temperature gradient where $\Delta T > 2^\circ\text{C}$ or $\Delta z > 10\text{ m}$ [Fiedler and Talley, 2006]. A low *N ratio* suggests a high relative abundance of *F. profunda* and consequently a deep thermocline (Figure 5). Conversely, a large *N* index indicates a low relative abundance of *F. profunda* and a shallow thermocline.

Six core top samples were removed from the dataset because coccolith abundances were either too low for accurate counting (< 20 individuals per field of view) or species known to have gone extinct prior to the Quaternary were identified in the sample, suggesting significant reworking of nannofossils. An additional three core-top samples recovered from the vicinity of the Galapagos Islands record lower than expected N ratios for their modern day thermocline depths relative to the rest of the dataset (Figure 5), and we therefore also remove these data from the calibration. The preservation of coccoliths in all core-tops samples is excellent as determined by the CEX' index. All samples exhibit high CEX' values above the critical value of 0.63 [Boeckel and Baumann, 2004].

5.4 Thermocline Depth Reconstructions

We reconstruct thermocline depth at sites ME0005A-15MC/17JC (4.62°N), GGC 43/CDH 41 (1.25°S), and VNTR01-13GC (3.09°S) over the past 30 kyrs using the established N ratio calibration (Figure 6). All locations exhibit a shallower thermocline during the LGM and deglaciation relative to present day. The thermocline depth at each site converges towards comparable values, between 20 and 30 m depth, during the last deglaciation and diverges at the onset of the Holocene towards present day. Site VNTR01-13GC exhibits the largest range in thermocline depth over the past 30 kyrs with a minimum depth of 20.9 m observed 15.5 ka and maximum values observed during the late Holocene. Because core-tops in the vicinity of the Galapagos fall off the calibration (Figure 5), the N ratio inferred thermocline depth may be offset from the true thermocline depth at the Galapagos site. However, we expect that the trend is accurately reconstructed.

6. Discussion

6.1 No evidence for ENSO-like changes in the EEP mean state

Deepening (shoaling) of the EEP thermocline is a characteristic feature of El Niño (La Niña) events and leads to a number of correlations between parameters that leave proxy traces in the sedimentary record. ENSO dynamics lead to a strong positive relationship between thermocline depth and SST and a strong negative relationship between primary production and thermocline depth (and SST) on interannual timescales [e.g. *Cane et al.*, 1997; *Barber et al.*, 1996]. Over the past 30 kyrs we find that these characteristic ENSO relationships between thermocline depth, SST, and productivity breakdown. Open-ocean sites (ME0005A-15MC/17JC and VNTR01-13GC) north and south of the equator exhibit no statistically significant relationship between thermocline depth and SST or productivity (Figure 7). Near the Galapagos Islands thermocline depth is weakly correlated with SST ($R^2=0.50$, $p<0.01$) and coccolithophorid production ($R^2=0.34$, $p<0.05$), but productivity exhibits a *positive* correlation with both parameters (SST: $R^2=0.35$, $p<0.01$), exactly the opposite expected from ENSO dynamics.

The relationship between SST and primary production was examined at three additional sites in the EEP, spanning 3.99°S to 4.61°N, and include both open-ocean and coastal upwelling sites. All sites exhibit statistically significant relationships between SST and productivity but not always the inverse correlation expected from ENSO-related dynamics (Figure 8a). Open-ocean sites in the EPWP and the southern edge of the ECT maintain the modern ENSO-like relationship between production and SST, with cool SSTs leading to greater coccolithophorid production ($R^2=0.72$, $p<0.01$ and $R^2=0.36$, $p<0.01$, respectively) and diatom productivity ($R^2=0.$, 0.74 , $p<0.01$ and $R^2=0.84$, $p<0.01$,

respectively). Sites closer to perennial upwelling zones, however, near the Galapagos and N. Peru Margin exhibit a more complicated history. During the LGM and deglaciation, coccolithophorid production and SSTs are not significantly correlated, while the Peru Margin site exhibits a significant negative relationship between the two parameters ($R^2=0.19$, $p<0.05$). At the onset of the Holocene 11.8 ka, both sites show a significant rise in the production when SSTs rise (Galapagos: $R^2=0.55$, $p<0.01$; N. Peru Margin: $R^2=0.47$, $p<0.01$), again the opposite relationship expected for ENSO forcing. Although, we cannot eliminate the possibility that diatoms responded differently to SST variations due to an absence of diatom biomarker record from these sites, given the tight correlation between brassicasterol concentrations and C_{37} -alkenones at sites VNTR01-13GC, ME0005A-15MC/JC [Chapter 5, Figure 7] and ODP site 1240 [Calvo *et al.*, 2011] this scenario seems unlikely.

Based on the relationships between SST, thermocline depth, and paleoproductivity we suggest that ENSO-like changes in the east-west tilt of the equatorial thermocline do not characterize the paleoceanography of the EEP on millennial timescales since at least the last glacial period. ENSO variability should be strongest in the cold tongue but, even sites in the heart of it, do not exhibit the characteristic ENSO-relationships between thermocline depth and SST or productivity. These results support model results of *DiNezio et al.*, [2011], which do not simulate clear El Niño or La Niña patterns of SST change at the LGM. They find that on centennial and longer timescales the Bjerknes Feedback, which drives ENSO variability, is weak. Instead, zonal advection, evaporation, and cloud cover influence SST locally, leading to patterns of SST that do not resemble the modern ENSO pattern [Ma *et al.*, 1996; DiNezio

et al., 2009]. Future work in the western equatorial Pacific is required to confirm the lack of ENSO-like dynamics in the tropical Pacific over the past 30 ka; variability in EEP SST, thermocline depth, and paleoproductivity reconstructed here is just one part of the E-W contrast that characterize ENSO dynamics. Nevertheless, observed relationships between parameters in the east caution against interpreting trends in EEP SST and paleoproductivity under the ENSO framework on millennial and longer timescales.

6.2.1 Extratropical controls on thermocline depth in the EEP

On decadal and longer timescales remote forcing from the mid- and high-latitudes is likely more effective in determining the depth of the EEP thermocline than locally forced, ENSO dynamics [*Andreasen et al.*, 2001; *Boccaletti et al.*, 2004]. Climate change in the high latitudes can drive long-lived shifts in the position of the Hadley cells, and consequently the ITCZ, through the atmospheric bridge, with direct consequences for wind patterns and upwelling in the EEP. In addition, the oceanic tunnel connects the EEP subsurface to the mid- and high-latitudes by transporting waters from the extratropics directly into the EEP subsurface [*Toggweiler et al.*, 1991; *Liu and Philander*, 1995; *Gu and Philander*, 1997; *Lu et al.*, 1998; *Andreasen et al.*, 2001; *Boccaletti et al.*, 2004]. Thermal and/or salinity anomalies sourced from these regions significantly alter stratification in the EEP [*Bova et al.*, 2015] and interact with the surface-driven ENSO cycle in intriguing ways.

The relative impact of the atmospheric bridge and the oceanic tunnel on the EEP thermocline can be diagnosed by evaluating thermocline depth trends along a meridional transect. North-south movement of the Hadley cells will lead to disparate thermocline depth trends north and south of the ITCZ; a southward migration of the ITCZ as

reconstructed for the LGM [*Peterson et al.*, 2000; *Koutavas et al.*, 2003; *Koutavas and Lynch-Stieglitz*, 2004] would lead to weaker winds over the ECT and reduced equatorial upwelling and stronger winds in the EPWP. In contrast, subsurface anomalies transported via the oceanic tunnel will drive a uniform shoaling or deepening of the thermocline across the EEP. Subsurface waters are derived from the same, predominantly southern hemisphere sources [*Toggweiler et al.*, 1991; *Qu et al.*, 2009; *Bostock et al.*, 2010] across the entire EEP. Subsurface waters were cooler in the LGM relative to today [*Bova et al.*, 2015] and should therefore promote greater stratification between surface and thermocline waters and a shallow EEP thermocline.

Our thermocline depth histories from three sites along a N-S transect, spanning 3.09°S to 4.61°N, in the EEP exhibit approximately synchronous changes in thermocline depth over the past 30 kyrs, and are consistent with the oceanic tunneling system exerting primary control on the EEP thermocline depth since the LGM. The thermocline shoals to the shallowest observed depths relative to today, by 20 to 30 m, between 21 and 13 ka and becomes progressively deeper at all sites towards present day, synchronous with a rise in SOIW temperatures during the Younger Dryas [*Bova et al.*, 2015]. In addition, we observe a disconnect between thermocline depth and surface ocean SSTs, which suggests change in the thermocline depth was driven from below. A regionally synchronous change in thermocline depth by about 20 m is supported by two previous reconstructions based on planktonic foraminiferal abundance transfer functions [*Andreasen and Ravelo*, 1997] and $\delta^{18}\text{O}$ gradients between surface and thermocline dwelling planktonic foraminifera [*Faul et al.*, 2000]. It is therefore likely that trends in the depth of the EEP

thermocline since the LGM were set by the temperature and salinity of waters delivered to the base of the equatorial thermocline from the extratropics.

Although trends in thermocline depth variability since the LGM are synchronous across the region, the N-S structure of thermocline depth was different at the LGM in the EEP relative to today. Today the thermocline is deeper north of the equator than south of the equator but thermocline depth at the LGM was uniform across the equator. A more southerly position of the ITCZ, as reconstructed for the LGM, could explain this pattern; southward migration of the ITCZ leads to weaker winds over the ECT and a deeper thermocline south of the equator, while stronger winds in the EPWP drive a shoaling of the thermocline north of the equator. Shifts in the mean position of the trades winds may therefore alter the N-S thermocline depth gradient simultaneous to regional changes in thermocline depth driven from the subsurface.

7. Concluding Remarks

Reconstruction of the change in thermocline depth, SST, and primary productivity at four sites across the eastern equatorial Pacific demonstrates that ENSO is not the dominant source of variability in the EEP during the LGM to Holocene transition. Correlations between SST, thermocline depth, and primary productivity do not exhibit the characteristic ENSO relationships observed today. Changes in mixed layer SST and productivity do not accompany changes in thermocline depth, perhaps indicating a greater influence of local wind forcing on the mixed layer and remote forcing on thermocline depth [*Fiedler and Talley, 2006*]. We therefore conclude that transmission of high latitude climate signals via the oceanic tunneling system exerts primary influence on the depth of the EEP thermocline over the timespan of our records. If we are correct,

inferring paleo-ENSO state from a single near-surface proxy, or the variance in a single near-surface proxy may be misleading.

The conclusions of this study have important implications for predicting the future response of the equatorial Pacific to impending climate change. Temperatures are projected to rise globally but polar amplification means that regions closer to the poles will warm fastest [Masson-Delmotte et al., 2006]. We therefore expect subsurface waters, sourced from the extratropics, to warm faster than those at the EEP surface, leading to a reduction in upper ocean stratification in response to anthropogenic climate change. Reduced stratification will in turn produce a mean deepening of the equatorial Pacific thermocline and reduced heat uptake in the tropical Pacific. The response of ENSO to such a change remains uncertain [DiNezio et al., 2009; Vecchi and Soden, 2007], but we can predict that a deeper EEP thermocline will lead to a reduction in CO₂ evasion from the EEP over the coming centuries [Le Borgne et al., 2002].

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Figures:

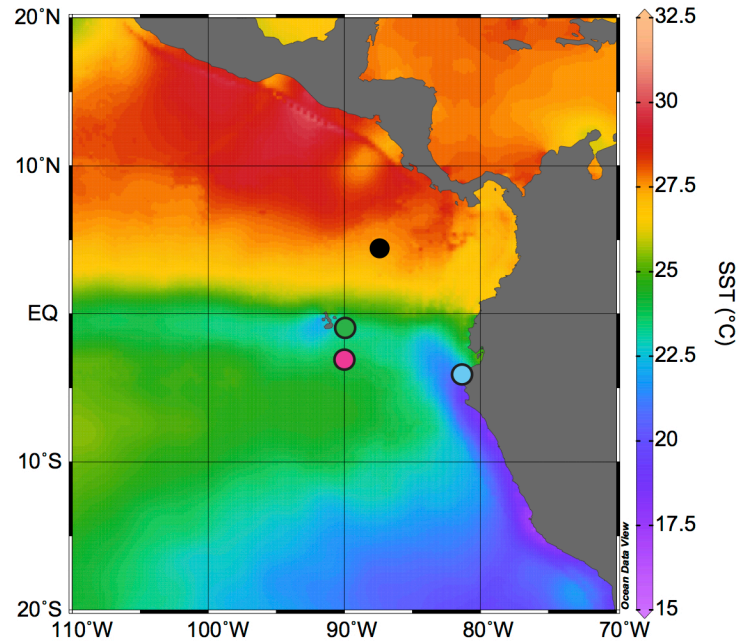


Figure 1. Locations of sediment cores used in this study overlaid on a map of mean annual surface ocean temperatures [Locarnini *et al.*, 2013]. Black circle: ME0005A-15MC/17JC, Green Circle: GGC 43/CDH 41, Pink circle: VNTR01-13GC, Blue circle: CDH 23/CDH 26. Map constructed using Ocean Data View software.

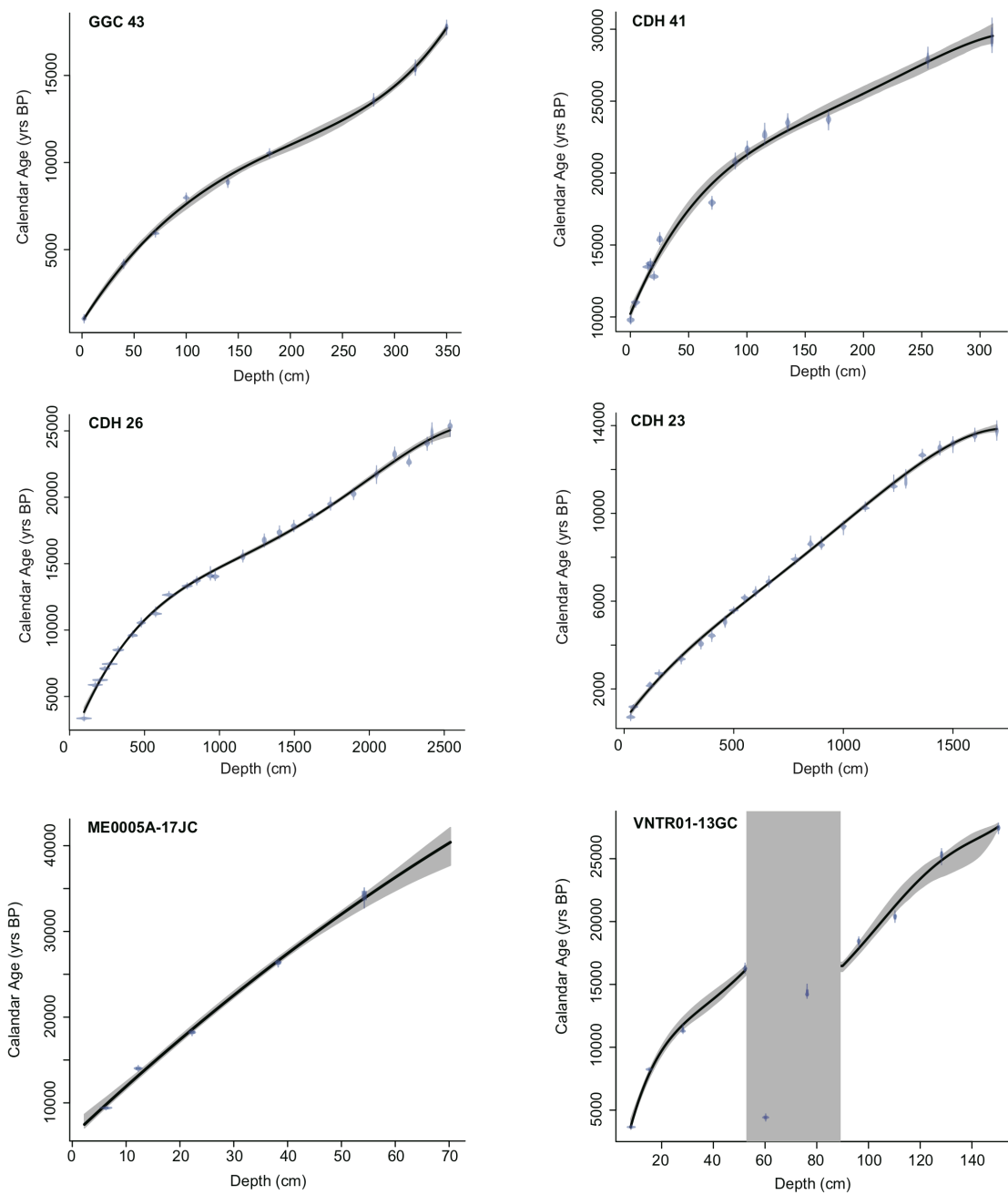


Figure 2. Age-depth models for sediment cores used in this study. All age models were constructed using the Clam age modeling software package for R [Blaauw, 2010]. Grey error envelopes represent one standard deviation.

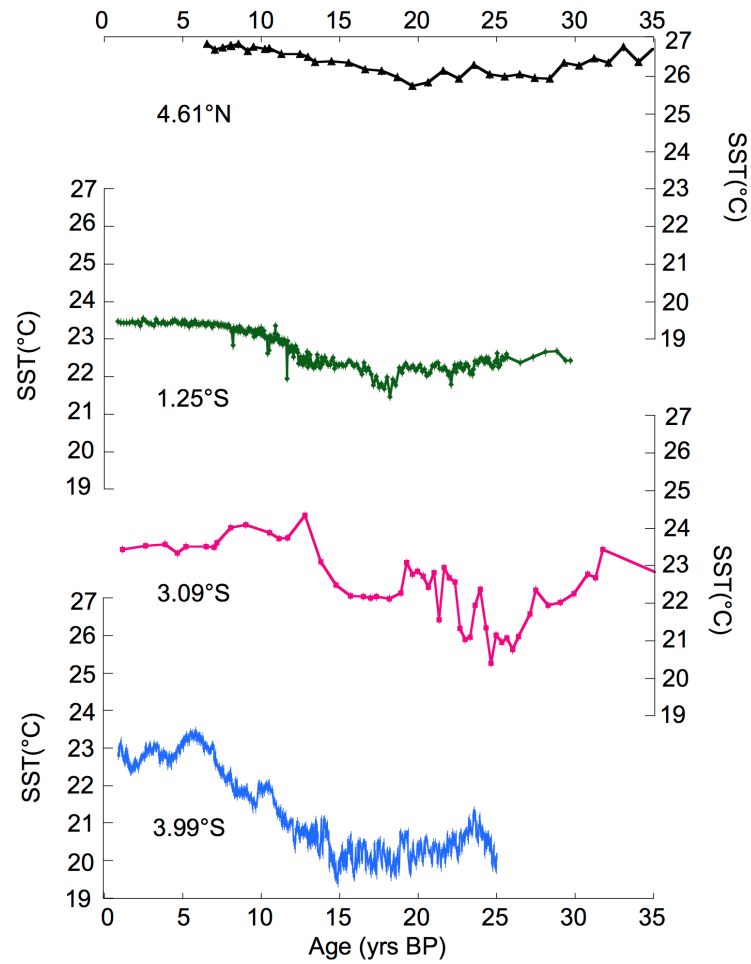


Figure 3. Alkenone SST reconstructions from four sites in the EEP. The magnitude of warming during the deglaciation (11.8 to 18 ka) is larger and delayed at the two southernmost locations (pink: VNTR01-13GC, blue: CDH 23/CDH 26) relative to the GU (green: GGC 43/CDH 41) and EPWP (black: ME0005A-15MC/17JC) sites.

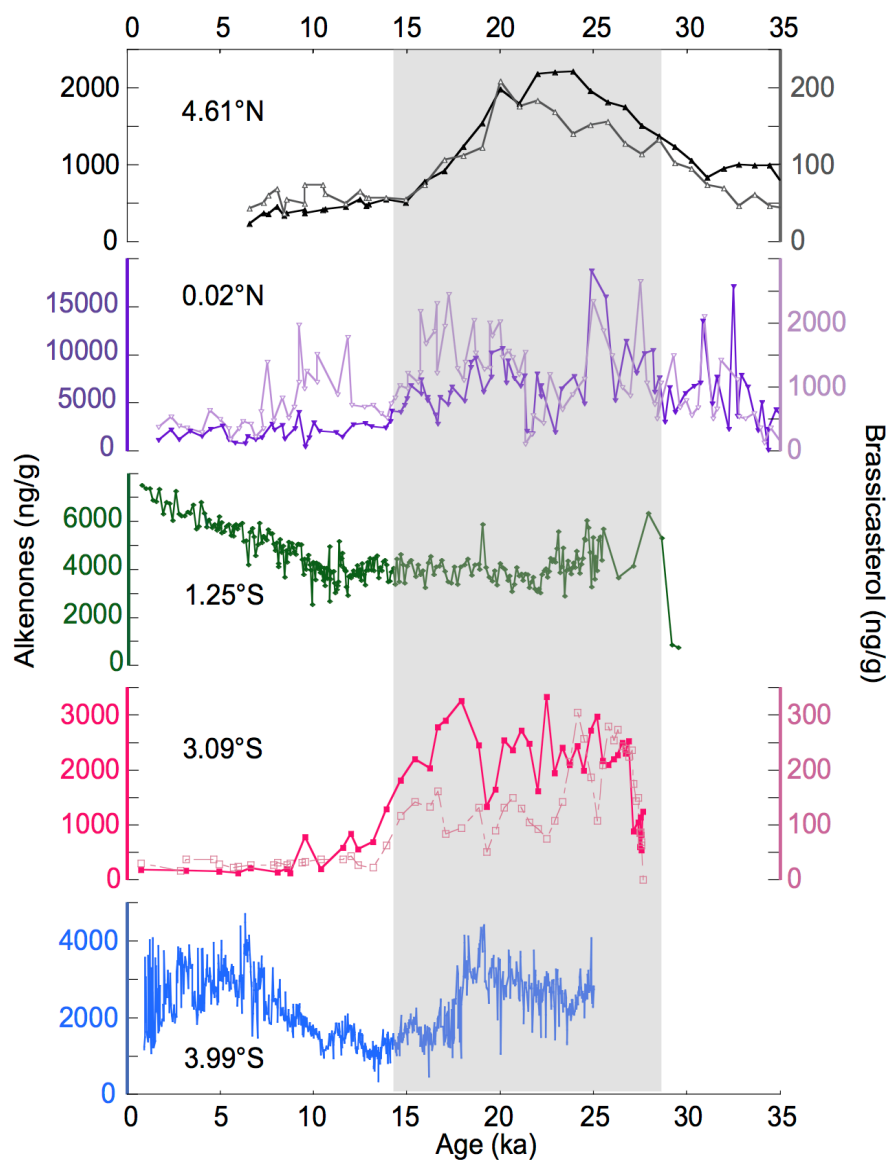


Figure 4. Organic biomarker abundances at five sites in the eastern equatorial Pacific spanning 4°S to 4.5°N, including a previously published record from ODP site 1240 (purple) [Calvo *et al.*, 2011]. Alkenone abundances (dark) reflect coccolithophorid productivity and brassicasterol concentrations (light) reflect diatom production rates.

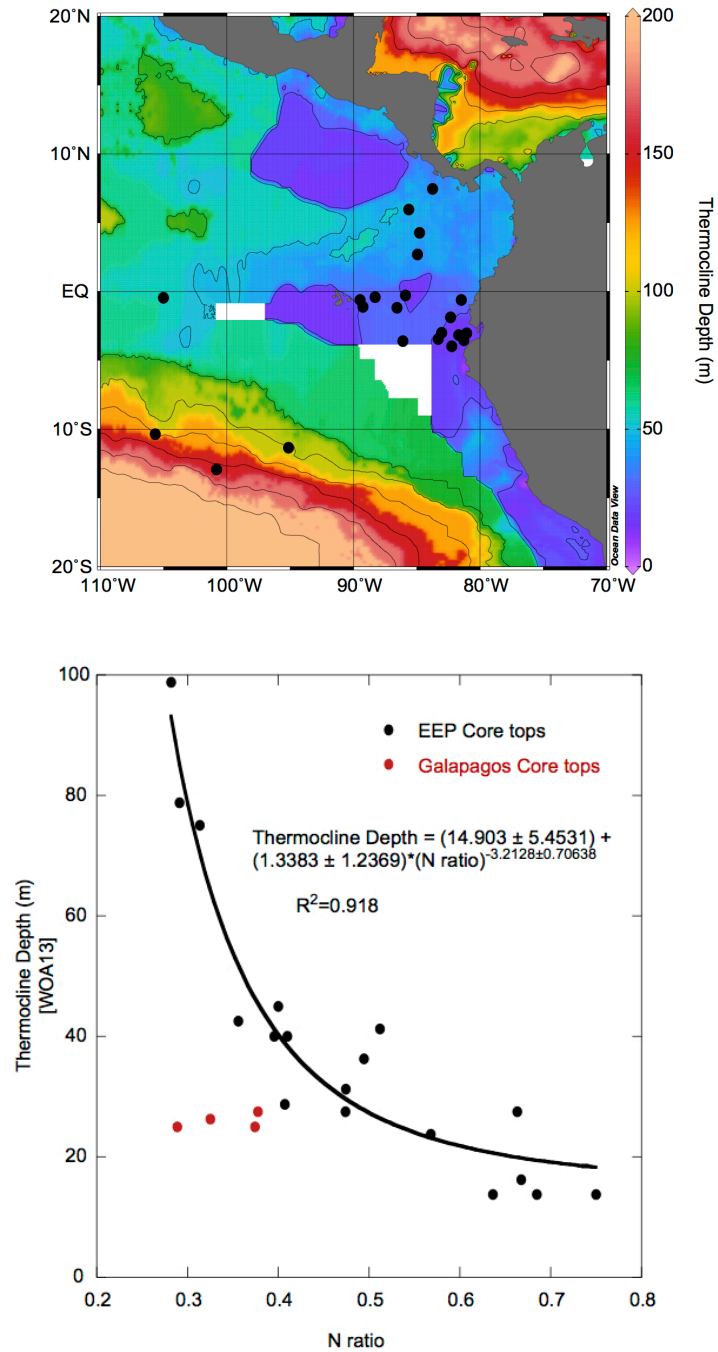


Figure 5. (top) Map of thermocline depth in the EEP between 20°S and 20°N with locations of multicore core top samples used in the *N ratio*-thermocline depth calibration plotted as black circles. White regions represent areas with low data coverage and thus bad estimates of thermocline depth. (bottom) Relationship between the *N ratio* of core

top samples and the modern thermocline depth at the core locations as estimated from the WOA 13 database [Locarini *et al.*, 2013]. We find a significant relationship between the *N ratio* of core top samples and observed thermocline depth ($R^2=0.918$, $p<0.01$). Samples collected near the Galapagos Islands (red) exhibit lower than expected *N ratios* given the modern thermocline depth at their locations. We exclude these data points from the calibration.

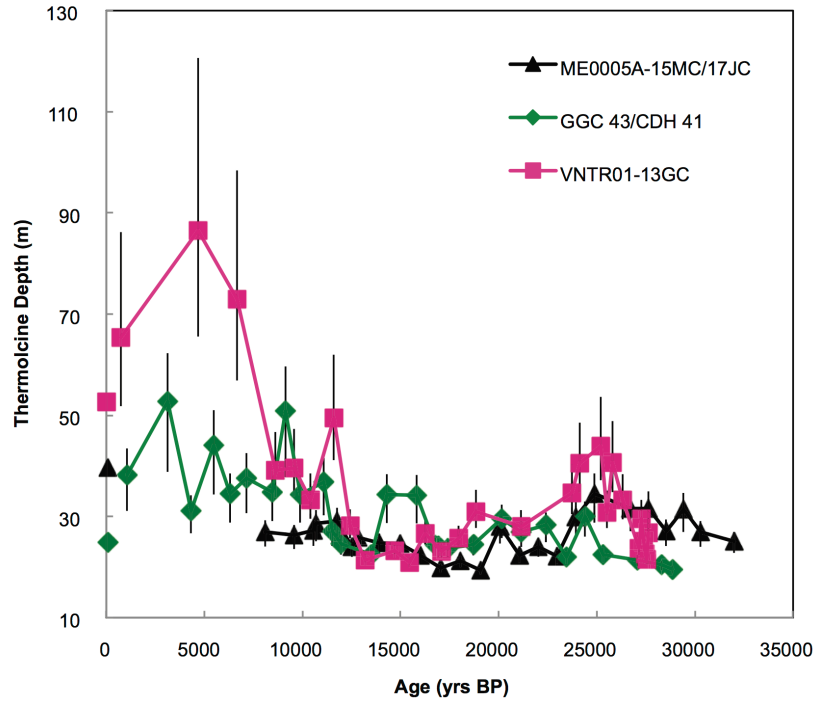


Figure 6. Thermocline depth histories at sites ME0005A-15MC/17JC (4.61°N), GGC 43/CDH 41 (1.25°S), and VNTR01-13GC (3.09°S). Thermocline depth is uniformly lower during the last glacial period and deglaciation and increases towards present day at all sites. Modern estimates, plotted on the y-axis, are from WOA13 database [Locarini et al., 2013]. Error bars represent one standard deviation.

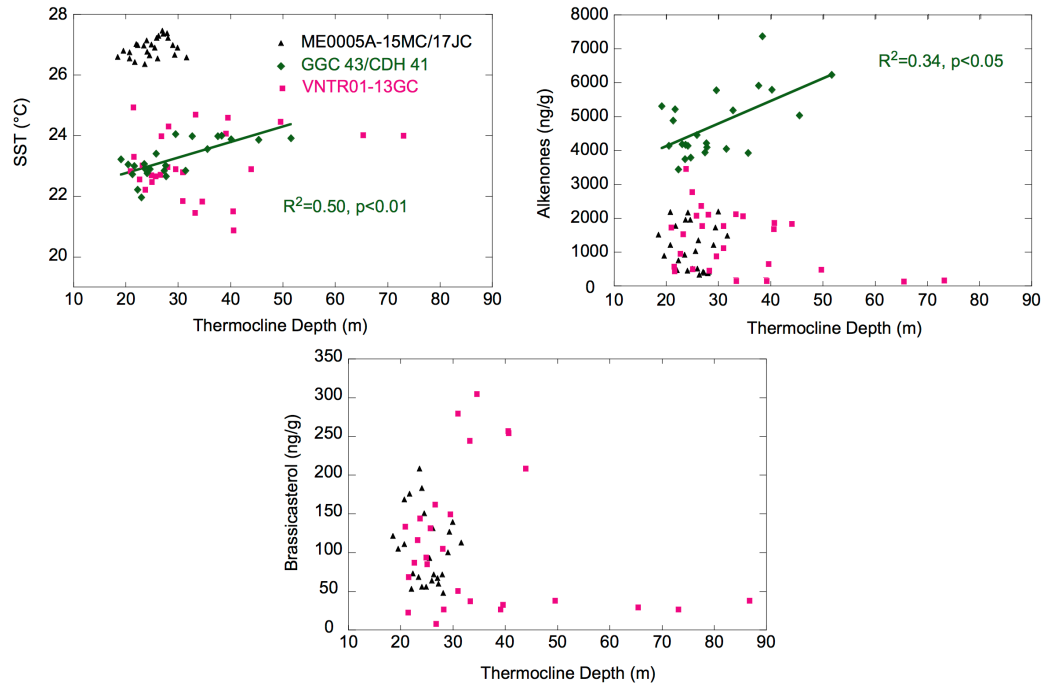


Figure 7. Relationships between thermocline depth and mixed layer properties (SST and phytoplankton production). Thermocline depth and SST at the Galapagos site (GGC 43/CDH 41) are significantly correlated. However, the relationship between thermocline depth and coccolithphorid productivity is the inverse of that expected from ENSO dynamics.

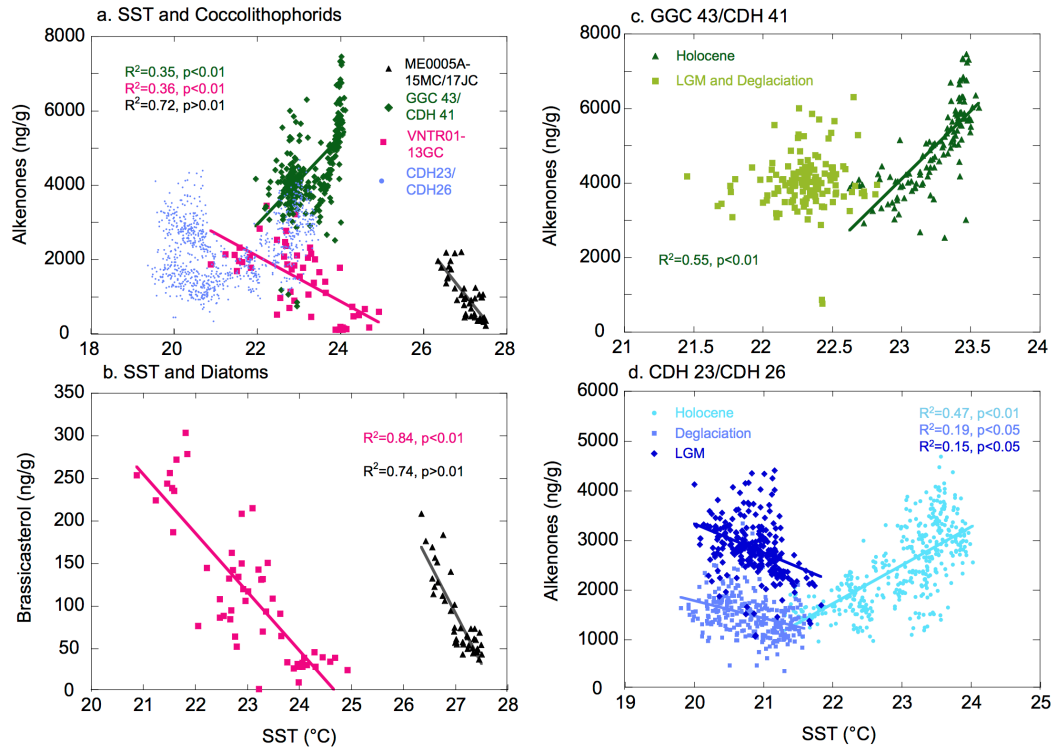


Figure 8. Relationships between SST and phytoplankton productivity at four sites in the EEP. Open-ocean sites exhibit the expected relationship between SST and coccolithophorid and diatom production (a, b), with cool temperatures leading to higher rates of primary production for both groups. (c,d) The relationship between SST and coccolithophorid production near the Galapagos and off the coast of northern Peru exhibit different relationships during the LGM, deglaciation, and Holocene (0-11.8 ka). Both sites exhibit a statistically significant positive relationship between SST and alkenone abundances during the Holocene. At the Peru Margin we observe the inverse relationship during the LGM (older than 18 ka) and deglaciation (11.8-18 ka). Within the GU there is no statistically significant relationship between productivity and SST during the last deglaciation and LGM.

Table 1. Detailed information on sediment cores used in this study.

Core ID	Latitude	Longitude	Water depth(m)	¹⁴ C Reservoir Age (yrs)	Average Sedimentation Rate (cm/kyr)
ME0005A-15MC/17MC	4.61°N	86.70°W	905	440	2
KNR 195-5 GGC 43	1.25°S	89.69°W	617	500	19
KNR 195-5 CDH 41	1.27°S	89.70°W	595	500	21
KNR 195-5 CDH 23	3.75°S	81.13°W	373	500	127
KNR 195-5 CDH 26	3.99°S	81.31°W	1026	500	133
VNTR01-13GC	3.09°S	90.82°W	3304	535	4.5

CHAPTER FIVE

No role for the eastern equatorial Pacific biological pump in atmospheric
CO₂ change since the last glacial period

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Abstract

Changes in the rates of phytoplankton production in the Equatorial Eastern Pacific (EEP) may play a critical role in modulating orbital-scale climate change, yet a clear picture of their productivity history remains elusive. Reconstructions of the change in productivity based on indicators of diatom (% opal, $\delta^{30}\text{Si}$, brassicasterol), coccolithophore (C_{37} -alkenones, % calcite), and total algal production (C_{org} , carotenoid flux, $\delta^{15}\text{N}$) during the Last Glacial Maximum (LGM, 21 ka) and last deglaciation (18-11 ka) are inconsistent. It is unclear whether the disagreement between records arises from strong spatial heterogeneity in paleoproductivity, differing responses of algal groups to oceanographic and climatic drivers, or proxy biases. In this study we reconstruct coccolithophorid (C_{37} -alkenones) and diatom paleoproductivity (brassicasterol) at multiple sites spread across the EEP. We find that total production increased during the last glacial period at open-ocean sites, likely due to aeolian iron fertilization, but we do not observe the expected shift to greater diatom relative to coccolithophorid productivity. Diatoms and coccolithophores change similarly during the last glacial period and deglaciation, indicating that shifts between calcareous and non-calcareous algal groups had a small impact on the biological carbon pump and, consequently, little effect on the evasion of CO_2 from the EEP. Disagreement among paleoproductivity records arises from a combination of proxy bias and true regional heterogeneity.

1. Introduction

Biological production changes in the eastern equatorial Pacific (EEP) are often implicated in mechanisms for glacial-interglacial variations in atmospheric carbon dioxide (CO_2) [e.g. *Matsumoto et al.*, 2002; 2008; *Cermeño et al.*, 2008; *Pichevin et al.*,

2009]. Today, the region is responsible for nearly two-thirds of the global marine CO₂ flux to the atmosphere each year, despite composing less than 0.1% of the surface ocean [Takahashi *et al.*, 2002]. High CO₂ fluxes are the result of physical and biological processes; easterly and alongshore winds drive upwelling along the equator and the Peru Margin bringing cool, nutrient- and CO₂-rich sub-thermocline waters to the surface ocean [Pennington *et al.*, 2006; Brzezinski *et al.*, 2011; Moore *et al.*, 2013]. Currently, silicate and iron limitations restrict complete carbon uptake by phytoplankton, but, under (micro)nutrient replete conditions, such as those hypothesized during the last glacial period, an increase in the efficiency of phytoplankton production may have reversed the role of the EEP in the global carbon cycle and driven a reduction in atmospheric CO₂ [e.g. Brzezinski *et al.*, 2002; Matsumoto *et al.*, 2002; 2008; Cermeño *et al.*, 2008; Pichevin *et al.*, 2009; Murray *et al.*, 2012].

Importantly, the strength of the biological carbon pump is determined not by the total production but the relative export of organic carbon versus calcium carbonate from the surface ocean [e.g. Kohfeld *et al.*, 2005]. The relative contribution of calcareous (coccolithophores) versus non-calcareous (diatoms) phytoplankton to export production alters surface ocean pCO₂ [e.g. Calvo *et al.*, 2011]. Diatoms and coccolithophores both take up CO₂ during photosynthesis but coccolithophores also produce a carbonate skeleton that releases CO₂ back into surface waters [e.g. Calvo *et al.*, 2011]. A decrease in coccolithophores relative to diatoms therefore reduces the efficiency of the carbonate pump causing a decrease in surface ocean pCO₂ and a reduction in atmospheric CO₂ for the same production of organic carbon. Evidence suggests the relative abundance of diatoms versus coccolithophores is regulated by the supply of biologically important

nutrients, particularly silicate and iron, to the surface ocean, because nutrient replete conditions favor diatoms at the expense of coccolithophores (e.g. *Matsumoto et al.*, 2002).

Constraining the change in calcareous and non-calcareous phytoplankton productivity at the LGM compared to today is thus the subject of a large body of recent paleoceanographic work [e.g. *Lyle et al.*, 1988; *Farrell et al.*, 1995; *Loubere et al.*, 2003; *Loubere et al.*, 2004; *Bradtmiller et al.*, 2006; *Koutavas and Sachs*, 2008; *Chazen et al.*, 2009; *Pichevin et al.*, 2009; *Arellano-Torres et al.*, 2011; *Dubois et al.*, 2010; *Calvo et al.*, 2011]. Published paleoproductivity records show regionally heterogeneous changes during the deglaciation and LGM (Figure 1) [*Loubere et al.*, 2003; *Loubere et al.*, 2004; *Bradtmiller et al.*, 2006; *Koutavas and Sachs*, 2008; *Pichevin et al.*, 2009; *Dubois et al.*, 2010; *Arellano-Torres et al.*, 2011; *Calvo et al.*, 2011], which makes it difficult to assess the role of the EEP in glacial-interglacial CO₂ change. We propose that by incorporating spatial analysis of LGM and deglacial paleoproductivity we can provide a framework for explaining the diversity of changes observed in the region and for identifying the mechanisms for paleoproductivity changes. In this study, we analyze organic biomarkers of diatoms and coccolithophores from multiple sites, spanning 4°S to 4.6°N, and begin to reconcile these seemingly conflicting productivity records from the EEP.

2. Productivity in the EEP

2.1. Spatial Gradients

Primary production in the EEP is a function of macronutrient and iron supply to the surface mixed layer [*Pennington et al.*, 2006]. Macronutrients (nitrate, phosphate, and silicate), sourced primarily from subsurface intermediate waters formed in the northern

and southern high latitudes, are delivered to the surface ocean fueling high rates of productivity in coastal and equatorial upwelling centers [Sarmiento *et al.*, 2004; Fiedler and Talley, 2006]. Despite the high rates of production, nitrate and phosphate are never fully depleted in EEP surface waters and the region is classified as a High-Nutrient Low-Chlorophyll region. It has been hypothesized that low iron concentrations limit growth [Brzezinski *et al.*, 2011], particularly in EEP open ocean locations where aeolian dust is the only significant source of iron [Behrenfeld and Kolber, 1999]. Iron is less limiting along the continental margin and the Galapagos Islands, due to interactions of currents with margin sediments, riverine inputs, and upwelling from the relatively iron-rich Equatorial Undercurrent (EUC) [Coale *et al.*, 1996; Gordon *et al.*, 1997; Johnson *et al.*, 1999]. Variation in upwelling strength and iron availability subsequently lead to strong regional variations in productivity across the region.

We identify four zones within the EEP based on physical oceanographic regimes and nutrient restrictions [Pennington *et al.*, 2006]. These include: (1) the Peruvian Coastal Upwelling (PCU), (2) the Galapagos Upwelling (GU), (3) the Equatorial Upwelling (EU), and (4) the eastern Pacific warm pool (EPWP) (Figure 2). The PCU is the most productive zone in the EEP, yielding nearly $1.3 \text{ kg C m}^{-3} \text{ yr}^{-1}$. Here alongshore winds drive surface waters away from the South American coast in the PCU system, bringing cool, macronutrient- and iron-rich subsurface waters from depth to the surface ocean [Pennington *et al.*, 2006]. Waters in the GU system are also highly productive, generating $0.7 \text{ kg C m}^{-3} \text{ yr}^{-1}$. Collision of the eastward flowing EUC with the islands drives perennial upwelling of modified subtropical mode water with contributions from southern ocean mode and intermediate waters (SOIW) [Kessler, 2006]. The EU zone and

the EPWP support lower levels of production, 0.23 and 0.18 kg C m⁻³ yr⁻¹, respectively, because macronutrients and iron are more limited [Pennington *et al.*, 2006]. Equatorial upwelling relies on easterly winds to drive divergence but, east of the Galapagos, the trades are primarily meridional. The EPWP lies beneath the ITCZ where low wind speeds, excess precipitation, and a seasonally large heat flux drive a strongly stratified surface layer [Kessler *et al.*, 1998; Kessler, 2006].

2.2 Temporal Variability

ENSO is the primary source of temporal variation in productivity over the historical record, larger even than the seasonal cycle [Chavez, 1995; Pennington *et al.*, 2006; Wang and Fiedler, 2006]. Its effects are felt most dramatically in the EU and coastal boundary regions ($\pm 1-2^{\circ}\text{C}$) and much less so in the EPWP [Fiedler and Talley, 2006]. ENSO drives strong and coherent changes in phytoplankton production. During an El Niño event, primary production rates and chlorophyll levels across the entire EEP decrease by several-fold. In coastal regions, the decline in phytoplankton abundance is accompanied by a shift in the community composition as well, with diatoms and other large-celled phytoplankton being replaced by picophytoplankton that are characteristic of the nutrient-limited open ocean [e.g. Chavez, 1989; Chavez *et al.*, 1991]. Although more work has been done to characterize the impacts of El Niño, strong La Niña events also impact the regional ecology and trigger large phytoplankton blooms within the EU system [Behrenfeld *et al.*, 2001].

3. Site Locations

In this study, we reconstruct the productivity history of each of the four key oceanographic zones of the EEP outlined above (Figure 2, 3). Core ME0005A-

15MC/17JC was recovered from the EPWP. Here, surface waters are sourced from the North Equatorial Countercurrent, which brings warm, nutrient poor surface waters to the core site [Kessler *et al.*, 1998; Kessler, 2006]. Cores GGC 43 and CDH 41 were recovered on the Galapagos Platform just north of Española Island and records surface water change in the GU system. Surface waters around the islands reflect the mixed influence of local upwelling around the islands and the eastward flowing South Equatorial Current [Kessler, 2006]. Evolution of the PCU system is characterized by cores CDH 23/CDH 26, retrieved from the Gulf of Guayaquil. Coastal upwelling, as well as competition between north and south flowing surface currents affects the core sites [Kessler, 2006; Chaigneau *et al.*, 2013]. Lastly, sites VNTR01-13GC together with previously published records from Calvo *et al.*, [2011] at ODP site 1240 record the evolution of the EU system in the EEP. ODP 1240 sits within the heart of the EU, while site VNTR01-13GC tracks its southern margin and is also under the influence of the South Equatorial Current, which advects Peruvian coastal waters offshore [Kessler, 2006; Chaigneau *et al.*, 2013].

4. Methods

4.1. Age models

Core chronologies were constructed by applying polynomial fits to sets of AMS ^{14}C dates on planktonic foraminifera using the Clam age model software [Blaauw 2010]. Radiocarbon measurements were made at the National Ocean Sciences Accelerator Mass Spectrometry facility at Woods Hole Oceanographic Institute. Planktonic foraminifera were picked from the $>150\ \mu\text{m}$ size fraction. All ^{14}C dates were converted to calendar ages using the marine13 radiocarbon age calibration curve [Reimer *et al.*, 2013]. We

assume constant reservoir ages at each of the sites based on the nearest modern measurements from the marine reservoir age database. For more details on the construction of the age-depth models used in this study see *Chapter 4*.

4.2 Biomarkers

We analyzed C₃₇-alkenones and brassicasterol concentrations at four new sites in the EEP: ME0005A-15MC/17JC, CDH 41/GGC 43, VNTR01-13GC, and CDH 23/CDH 26 (Figure 4). Alkenones are straight-chain hydrocarbons produced by the haptophyte algae *Emiliana huxleyi* and *Gephyrocapsa oceanica*, which represent the two dominant surface dwelling coccolithophores in the EEP [Okada and Honjo, 1973]. The concentration of C₃₇-alkenones in the sediment tracks coccolithophorid production [e.g. Lee and Schneider, 2005] and perhaps even total production, including siliceous diatoms, as demonstrated by Bolton *et al.* [2010] in the late Pliocene EEP. Brassicasterol concentrations are used to trace diatom production rates [e.g. Higginson *et al.*, 2004; Calvo *et al.*, 2011]. Brassicasterol is the major sterol produced by some diatoms, but may also be generated by haptophyte algae and cryptophytes [Volkman, 1986; Rampen *et al.*, 2010]. Therefore, the ratio of brassicasterol to alkenone concentrations must be used to evaluate shifts between coccolithophore versus diatom dominated production regimes. Deposition and preservation of alkenone and sterol biomarkers depend on a number of factors including zooplankton herbivory, sediment accumulation, and water column oxicity [Herbert *et al.*, 2003]. Sterols are more labile than alkenones and are observed to decrease in relative abundance in the water column and into the sediments [Wakeham *et al.*, 1997].

One to two grams of freeze-dried sediment was extracted in a Dionex 200 Accelerated Solvent Extractor with dichloromethane. Samples from sites CDH 26, CDH 23, GGC 43 and CDH 41 were analyzed only for alkenone biomarkers. After extraction samples were evaporated to dryness under a stream of nitrogen gas and spiked with an internal standard before analysis on a Gas Chromatograph with a Flame Ionization Detector (GC-FID). Alkenone records from the northern Peru Margin sites (CDH 23 and CDH 26) were published previously by *Bova et al.*, [2015] and provide a measurement approximately every 35 years.

Samples from sites VNTR01-13GC and ME0005A-15MC/17JC were analyzed for both alkenones and brassicasterol (24-methylcholesta-5,22-dien-3 β -ol) concentrations. After extraction these samples were reacted in 2 N potassium hydroxide in methanol and water (95:5) in order to hydrolyze esters and reduce interferences during gas chromatograph analysis. Samples were then further separated using silica gel chromatography with hexane, dichloromethane, and methanol as eluents. The methanol fraction was derivatized with bis(trimethylsilyl)trifluoroacetamide. The methanol and DCM fractions were then evaporated under a stream of nitrogen gas and reconstituted in toluene containing known concentrations of two internal alkane standards (*n*-hexatriacontane and *n*-heptatriacontane) before analysis on a GC-FID. Brassicasterol was not measured in cores from the Peru Margin and the Galapagos.

Alkenones in the DCM fraction were analyzed using the method of *Herbert et al.*, [1998]. Lab analytical error is equivalent to $\pm 0.1^{\circ}\text{C}$ based on 55 replicate analyses of a laboratory standard. Approximately 10% of our analyzed samples were run in duplicate with an average $U^{K'_{37}}$ -based temperature reproducibility of $\pm 0.05^{\circ}\text{C}$. Brassicasterol was

analyzed on an HP-1 column with an oven temperature programmed from 60°C (holding time of 0.8 minutes) to 180°C at 40°C/min, 180°C to 220°C at 10°C/min, 220°C to 315°C at 2°C/min, and 315°C to 325°C at 10°C/min. The brassicasterol compound was identified by analysis of a select number of samples on a gas chromatograph mass spectrometer. Analysis of synthetic brassicasterol confirmed peak identification. Thirty-nine brassicasterol samples were run in duplicate with an average reproducibility of ± 6.5 ng/g.

5. Results

5.1 Paleoproductivity estimates

Productivity records from across the EEP are regionally heterogeneous. Open ocean sites (ME0005A-15MC/17JC and VNTR01-13GC) exhibit maxima in both C₃₇-alkenone and brassicasterol abundances at the end of the last glacial period between approximately 15 and 31 ka, with a subsequent decline towards present day. The ratio of brassicasterol to alkenones remains less than 0.25 for the duration of both records, comparable to background levels at site ODP 1240 previously published by *Calvo et al.*, 2011. Both sites, however, exhibit maximum brassicasterol to alkenone ratios between 12 and 8 ka, synchronous with maximum values in the record from ODP 1240.

The Galapagos and Peru Margin sites exhibit markedly different productivity records than those from open ocean regions of the EEP (Figure 5). The Peru Margin site exhibits high alkenone abundances during the last glacial period, similar to the open ocean sites, but increases rather than decreases at the onset of the Holocene until 6.4 ka. Additionally, alkenone abundances are highly variable during the Holocene. At the Galapagos, alkenone abundances decrease between 27.9 and 22.0 ka from 6300 to 3200

ng/g and remain stable at 3900 ng/g from 22 to 11 ka. After 11 ka, alkenone abundances increase towards present day.

6. Discussion

The modern EEP is characterized by sharp gradients in macro- and micro-nutrient availability determined largely by the wind-driven circulation and the proximity of coastal upwelling zones [Pennington *et al.*, 2006]. Macro- and key micro-nutrients, such as iron, are supplied to the surface ocean via wind-driven upwelling of subsurface waters from beneath the thermocline. Iron is also delivered via aeolian dust deposition, but the supply is orders of magnitude lower than the subsurface supply [Coale *et al.*, 1996; Gordon *et al.*, 1997]. As a result, the highest rates of production in the modern EEP are centered in zones of strong coastal upwelling, such as the Peru Margin and the Galapagos (Figure 2) [Pennington *et al.*, 2006]. Because productivity is so variable across the region today it should come as no surprise that glacial to Holocene changes in productivity are not uniform across the region (Figures 4, 5). Here, we demonstrate a dichotomy in paleoproductivity trends between open-ocean sites, more than 100 km offshore, and coastal upwelling systems.

6.1. Iron limitations in the open-ocean EEP and the efficiency of the biological carbon pump

Paleoproductivity records from the EPWP (ME0005A-15MC/17JC) and within the EU system (VNTR01-13GC and ODP site 1240), all more than 100 km from the nearest coastline, exhibit maxima in diatom and coccolithophorid production during the last glacial period and early deglaciation between 15 and 33 ka (Figure 4). The near uniform response across a north to south transect (3°S to 4.6°N) in the open-ocean EEP

limits the possible causes of productivity change. We reject north-south competition of intermediate water masses and north-south shifts in wind-driven upwelling; both mechanisms should produce a north-south contrast in nutrient-availability, which we do not observe [Dubois *et al.*, 2010; Kienast *et al.*, 2006; Sarmiento *et al.*, 2004]. Instead, a change in the macronutrient content of upwelled waters sourced from the Southern Ocean and/or variation in iron availability are possible explanations.

Iron fertilization experiments demonstrate severe iron limitations in the modern open-ocean EEP [Hutchins and Bruland, 1998; Landry *et al.*, 2000; Pennington *et al.*, 2006]. After iron additions, the phytoplankton standing stock increases dramatically, with diatoms composing the majority of the bloom biomass [Hutchins and Bruland, 1998; Landry *et al.*, 2000; Barber and Hiscock, 2006; Pennington *et al.*, 2006]. The LGM represents a natural iron fertilization experiment. Aeolian iron inputs to the EEP increased 2 to 3-fold during the last glacial period in response to stronger glacial winds, a less robust hydrological cycle, and the expansion of desert source regions [Winckler *et al.*, 2008]. Productivity records presented here are consistent with a rise in total production during the last glacial period. Organic biomarker concentrations recorded at open ocean sites ME0005-15MC/17JC, VNTR01-13GC, and ODP 1240 suggest production of both coccolithophores and diatoms reached a maximum during the last glacial period and early deglaciation synchronous with an increase in dust delivery to the EEP, 33 to 15 ka [Winckler *et al.*, 2008; Calvo *et al.*, 2011] (Figure 4).

The tight correlation between C₃₇-alkenone and brassicasterol abundances (ME0005A-15MC/17JC: R²=0.83, p<0.01, VNTR01-13GC: R²=0.38, p<0.01) demonstrates that diatoms and coccolithophores responded equivalently to iron additions,

leading to no significant change in the strength of the biological carbon pump and CO₂ evasion in the EPWP and the southern margin of the EU system (Figure 6). This observation is in opposition to bottle experiments that show an increase in diatoms relative to coccolithophores when iron is more available [*Hutchins and Bruland, 1998*]. However, observations from the IronEx II mesoscale experiment suggest that, though iron additions do lead to greater diatom productivity, they do *not* lead to an increase in organic matter export relative to carbonate export. Efficient grazing of diatoms by protists recycles carbon in the surface mixed layer and maintains a low carbon export ratio [Landry et al., 2000]. Thus, despite greater iron inputs to the EEP during the LGM, the EEP biological carbon pump did not strengthen and therefore did not aid in glacial atmospheric CO₂ drawdown.

The early Holocene is instead the most likely interval for the EEP biological pump to sequester CO₂, despite no apparent rise in iron availability [*Winckler et al., 2008; Calvo et al., 2011*]. Closer examination of the ratio of brassicasterol to C₃₇-alkenones at sites ME0005A-15MC/17JC, VNTR01-13GC and ODP site 1240 [*Calvo et al., 2011*] reveals a coherent shift to enhanced diatom relative to coccolithophorid production between 12 and 8 ka (Figure 7). During this interval, silica leakage from the Southern Ocean at the end of the last deglaciation may have fueled the rise in diatom production [*Anderson et al., 2009; Calvo et al., 2011*]. Opal records from the Southern Ocean suggest surface waters around Antarctica, which feed the EEP upwelling systems, were more silica rich at this time [*Anderson et al., 2009*]. Although the relative increase at sites ME0005A-15MC/17JC and VNTR01-13GC is small, it is consistent with the much larger shift observed within the heart of the EU system at ODP site 1240.

Atmospheric CO₂ decreased by 10 ppmv between 10.4 and 8 ka [*Petit et al.*, 1999; *Monnin et al.*, 2001] synchronous with the rise in the ratio of diatoms to coccolithophores and thus the strength of the EEP biological carbon pump (Figure 7).

6.3 Macro-nutrient limitations in the Peru and Galapagos Upwelling Systems

In contrast to the open-ocean sites, phytoplankton living along the northern Peru Margin and the Galapagos Islands do not respond to changes in aeolian iron delivery over the past 30 kyrs (Figure 5). Paleoproduction rates in these locations instead reflect variations in macronutrient delivery via upwelling, zonal advection and/or variation in intermediate water nutrient content. Furthermore, because productivity in the Galapagos and Peru Margin upwelling systems has evolved differently, the nutrient supply at each site is likely controlled by different forcing mechanisms.

Upwelling immediately to the west of the Galapagos results from the collision of the eastward flowing EUC with the islands and is a perennial feature of the EEP [*Fiedler and Talley*, 2006; *Eden and Timmermann*, 2004]. Compared to the rest of the EEP, we posit that delivery of subsurface waters to the surface ocean was relatively stable at this location through time. We therefore use the record of GU productivity as an indicator of intermediate water nutrient content and infer no change in source water nutrient content until 11 ka, when coccolithophorid productivity increased at the site (Figure 5). The increase in productivity after 11 ka likely reflects greater nutrient concentrations in EEP sub-thermocline waters due to nutrient leakage from the Southern Ocean [*Anderson et al.*, 2009]. This is consistent with the observed rise in diatom to coccolithophorid productivity 12 ka at three open-ocean sites (Figure 7). Enhanced upwelling of sub-thermocline waters is not a good explanation for the observed increase in productivity

because the mean thermocline depth in the GU, the EPWP and EU increased beginning 13 ka and continued to increase towards present day [*Chapter 4*].

Paleoproductivity in the PCU system is influenced by a combination of source water nutrient change and local changes in the strength of wind-driven upwelling. Seasonal and interannual changes in the strength and direction of alongshore winds drive significant variations in regional productivity today [Pennington et al., 2006]. The paleoproductivity record from this site exhibits two major shifts over the past 30 ka: (1) at 18 ka, productivity decreases and (2) at 11 ka, productivity increases. The second shift aligns well with the rise in productivity observed in the GU system and an increase in diatoms relative to coccolithophores in the open-ocean EEP. Because the Galapagos and Peru Margin upwelling systems are fed by the same intermediate water sources, this change likely reflects the same increase in source water nutrient content. However, the first shift, 18 ka, does not align with any significant shift in productivity in the GU system. Thus, reduced production likely results from a decrease in local wind-driven upwelling and weaker along-shore winds. A significant negative relationship between SST and coccolithophorid production supports the role of local upwelling as it brings both cool and nutrient-rich waters to the surface ocean prior to 11 ka [*Chapter 4*]. We do not observe a similar relationship at the Galapagos.

6.2 Proxy Biases

We began this paper by evaluating a map of LGM productivity anomalies that showed a regionally heterogeneous response to glacial boundary conditions, unlike that observed today (Figure 1). We determined that phytoplankton in open-ocean versus coastal margin sites experienced different nutrient conditions, which led to diverse

productivity trends over the past 30 ka. These differences explain some, but not all of the observed differences in paleoproduction records from the EEP. Inorganic proxy records of paleoproduction in the open-ocean EEP based on the accumulation of calcite or opal [Loubere *et al.*, 2004; Bradtmiller *et al.*, 2006; Kienast *et al.*, 2006] and benthic foraminiferal assemblages in the sediments [Loubere *et al.*, 2003] are still at odds with organic biomarker and isotopic proxy records of production and nutrient utilization [Dubois *et al.*, 2010; Pichevin *et al.*, 2009; Calvo *et al.*, 2011]. ^{230}Th -normalized calcite and opal accumulation rates are 30 to 50% lower during the last glacial period relative to today and benthic transfer functions from the cold tongue indicate a 10 to 45% reduction in productivity [Loubere *et al.*, 2004; Bradtmiller *et al.*, 2006].

Preservation of proxy indicators may bias paleoproduction records. For example, the accumulation of calcite and opal in the sediment is sensitive to production, but also to dissolution during descent and preservation in the sediments. Opal is particularly susceptible to dissolution, as it is everywhere undersaturated with respect to seawater. Dubois *et al.* [2010] re-evaluated Th-normalized opal flux records from the EEP and found that preservation was, in fact, more important than production. Of course, organic biomarkers are also subject to preservational biases [Wakeham *et al.*, 1997; Herbert *et al.*, 2003]. However, they are typically well-preserved in low-oxygen environments such as the EEP [Herbert *et al.*, 2003] and also compare well to isotopic measures of nutrient utilization over the past 30 ka [Pichevin *et al.*, 2009]. We therefore suggest that biomarker records provide more reliable records of paleoproduction in the EEP over the past 30 ka than inorganic indices. Aside from paleoproduction records within coastal upwelling zones, once we remove inorganic proxy records from the map in Figure 1 we

find that productivity records consistently record greater total production at the LGM relative to the Holocene and today (Figure 8).

Concluding Remarks

Untangling the role of spatial variability, proxy bias, and ecological shifts in EEP paleoproductivity records has been a long-standing challenge for paleo-oceanographers. Paired diatom and coccolithophore biomarker records from multiple sites across the EEP suggest that differences between previously published proxy records reflect a combination of inter-proxy differences and spatial heterogeneity. Inorganic proxies are subject to dissolution and preservation biases in the sediment and confound regional interpretation of productivity change [Pichevin *et al.*, 2009; Dubois *et al.*, 2010]. Once removed, we find a regionally coherent productivity change in the LGM EEP. Total production, including coccolithophores and diatoms, increased during the LGM in the open-ocean in response to greater dust fluxes, while production within coastal upwelling regions did not (Figure 8). North-south shifts in upwelling zones and north-south competition between subsurface intermediate water masses were not the main drivers of paleoproductivity change in the open-ocean EEP, but may influence production in the comparatively iron-rich coastal upwelling zones.

Our records further suggest that the EEP was not a sink for atmospheric CO₂ during the last glacial period. CO₂ evasion depends on the export ratio of calcareous (coccolithophore) to non-calcareous (diatom) phytoplankton; an increase in diatom abundances relative to coccolithophores will drive a decrease in carbon outgassing from the EEP and perhaps turn the region into a carbon sink. Despite experimental evidence for an increase in diatoms relative to coccolithophores upon iron addition [Landry *et al.*,

2000; *Brzezinski et al.*, 2011], biomarker records indicate that export of alkenone-producing coccolithophores from the surface mixed layer increased in phase with diatoms during the last glacial period despite greater iron availability. Our results therefore indicate that iron-fertilization of the oceans will *not* be an effective method for mitigating the ongoing rise in atmospheric CO₂.

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Figures:

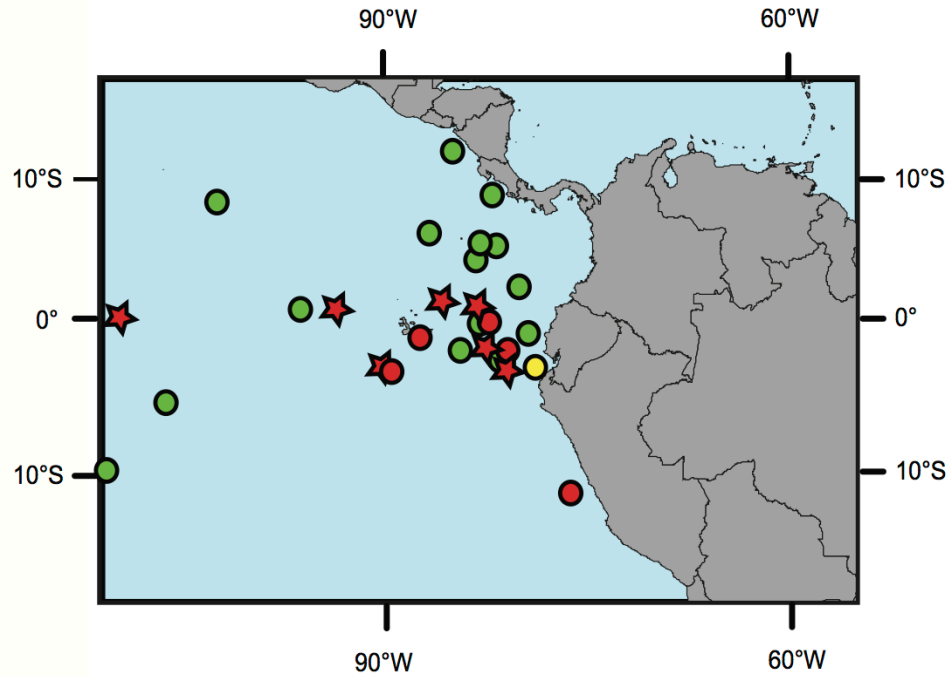


Figure 1. Locations of records of past productivity in the EEP spanning at least the past 25 ka. Red symbols indicate a decrease in productivity in the LGM relative to Holocene conditions. Green symbols indicate an increase in productivity in the LGM relative to Holocene conditions. Yellow symbols indicate no difference. Paleoproductivity proxies included in this figure include: the C_{37} -alkenones [Koutavas and Sachs, 2008; Calvo, *et al.*, 2011; this study], brassicasterol (diatom biomarker) [Calvo *et al.*, 2011; this study], ^{230}Th normalized opal and calcite fluxes [Loubere *et al.*, 2004; Bradtmiller *et al.*, 2006], opal mass accumulation rates [Arellano-Torres *et al.*, 2011], benthic foraminifera assemblages [Loubere *et al.*, 2003], diatom assemblages [Romero *et al.*, 2011], silicon isotopes [Pichevin *et al.*, 2009], and nitrogen isotopes [Pichevin *et al.*, 2009]. Stars indicate locations that use opal or carbonate fluxes to infer paleoproduction. Re-

evaluation of these proxy records suggests they reflect preservation rather than production [*Pichevin et al.*, 2009; *Dubois et al.*, 2010].

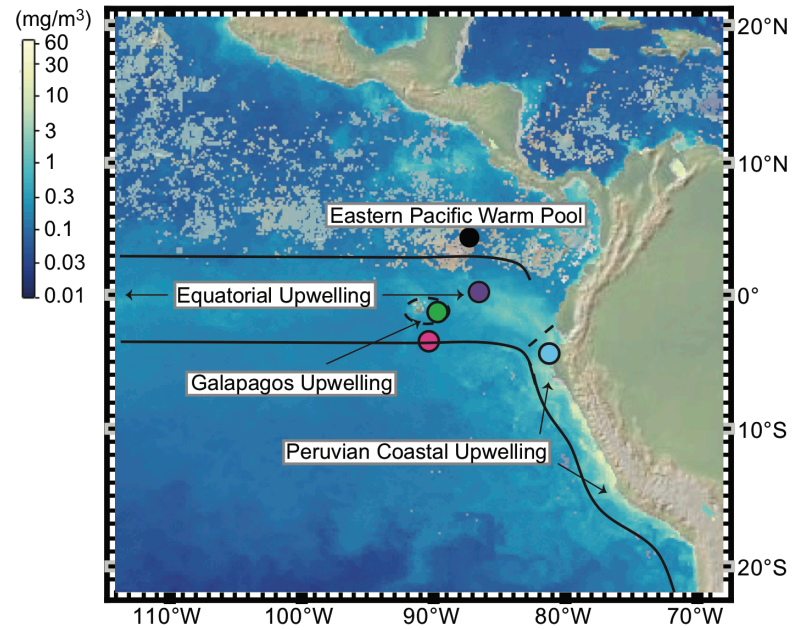


Figure 2. Map of mean annual chlorophyll concentrations in EEP surface waters (1 month – Aqua/MODIS). Black lines separate the region into four dynamically distinct oceanographic zones [Pennington *et al.*, 2006]. Filled circles indicate locations of sediment cores used in this study. Black circle: ME0005A-15MC/17JC, Green Circle: GGC 43/CDH 41, Pink circle: VNTR01-13GC, Blue circle: CDH 23/CDH 26. Purple circle: ODP 1240 Map constructed using the GeoMapApp software package.

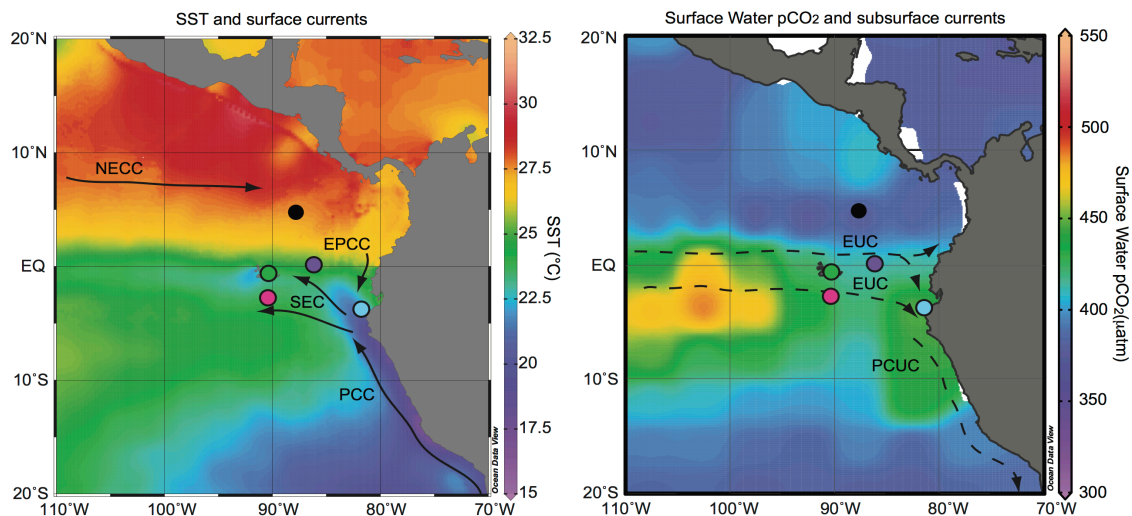


Figure 3. (left) Map of mean annual sea surface temperature [WOA13, *Locarnini et al.*, 2013] in the EEP overlaid with the major surface currents that affect our study sites [*Chaigneau et al.*, 2013]. (right) Map of the surface water partial pressure of CO₂ in the EEP [Takahashi et al., 2016] overlaid with the major subsurface currents [*Chaigneau et al.*, 2013]. Core sites are shown as filled circles. Black circle: ME0005A-15MC/17JC, Green Circle: GGC 43/CDH 41, Pink circle: VNTR01-13GC, Blue circle: CDH 23/CDH 26. Purple circle: ODP 1240 [Calvo et al., 2011]. Maps constructed using Ocean Data View software. EPCC: Ecuador-Peru Countercurrent; SEC: South Equatorial Current; NECC: North Equatorial Countercurrent; PCC: Peru Coastal Current; EUC: Equatorial Undercurrent; PCUC: Peru-Chile Undercurrent

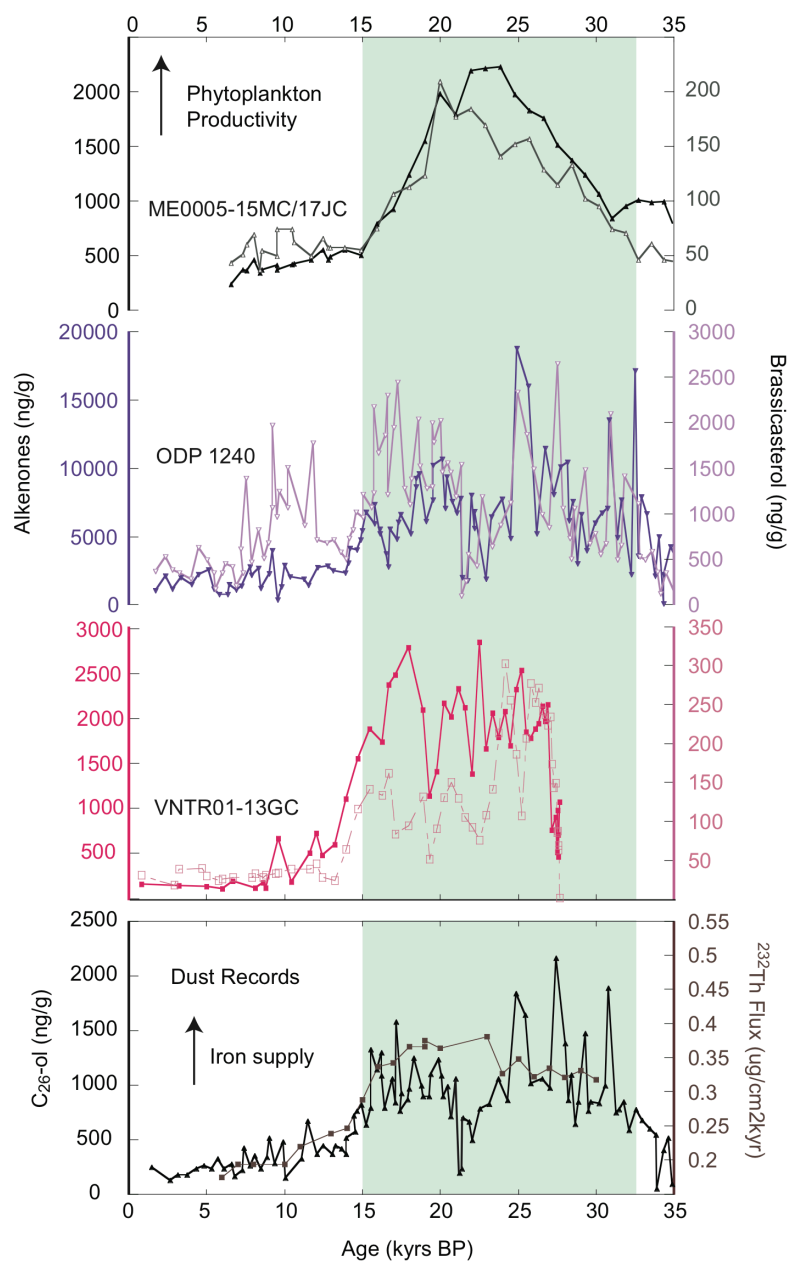


Figure 4. (top three panels) Organic biomarker abundances at three open-ocean sites in the EEP spanning 3.1°S to 4.6°N, including a previously published record from ODP site 1240 (purple) [Calvo *et al.*, 2011]. Alkenone abundances (dark) reflect coccolithophorid productivity and brassicasterol concentrations (light) reflect diatom production rates.

Black: ME0005-15MC/17JC; Pink: VNTR01-13GC (bottom panel) EEP dust records from the ODP site 1240 (C₂₆-ol) [Calvo et al., 2011] and nearby site TTN013-PC72 (²³²Th flux) [Winckler et al., 2008]. Dust-borne iron inputs are several-fold higher during the last glacial period and may be the cause of increased productivity in the open-ocean EEP between 15 and 33 ka.

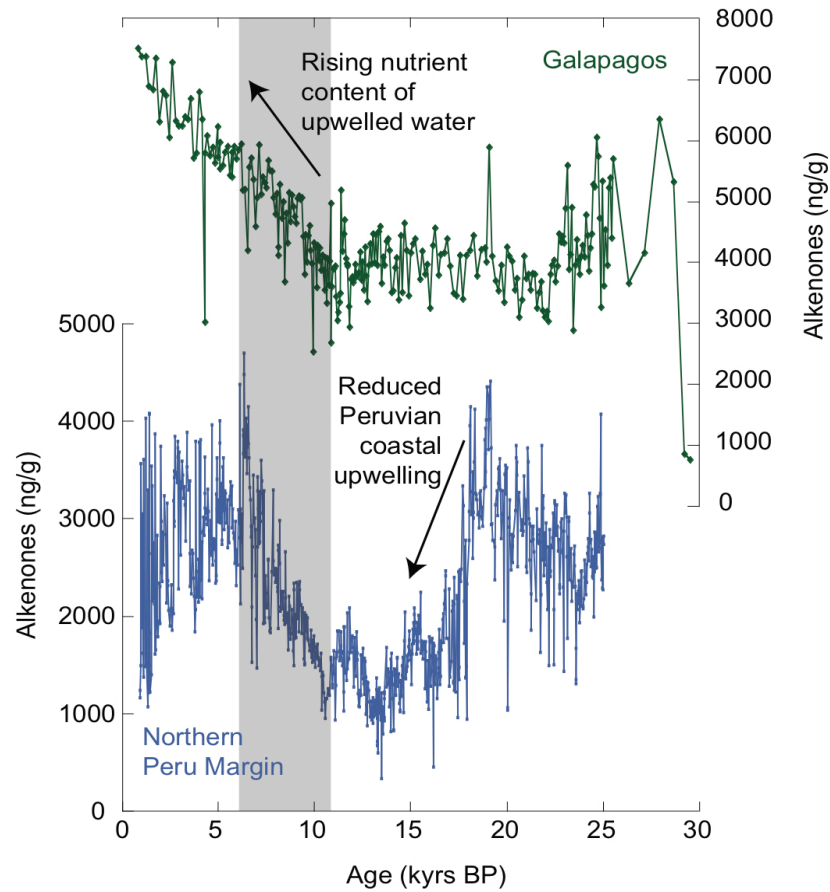


Figure 5. C_{37} -alkenone abundances, a proxy for coccolithohorid productivity, at the Galapagos (top, green) and the Peru Margin (bottom, blue) study sites. Coccolithophorid production increases at the onset of the Holocene in both records, indicating an increase in the nutrient content of the source water for both upwelling systems. The earlier decrease in production 18 ka along the Peru Margin, however, cannot be explained by a change in source water nutrients and, instead, likely reflects a reduction in wind-driven upwelling at the site.

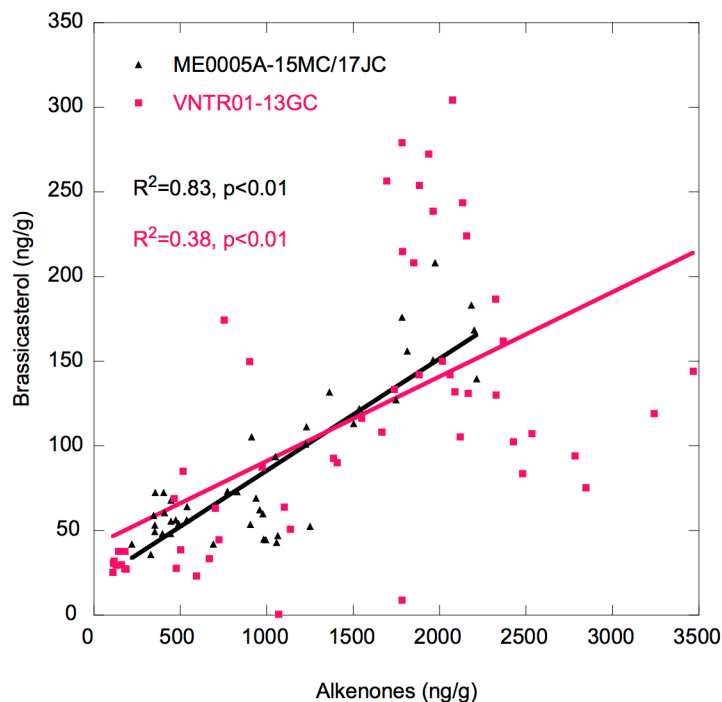


Figure 6. Relationship between coccolithophorid (alkenones) and diatom (brassicasterol) productivity at two sites north and south of the equator in the EEP. We observe a significant relationship between these two parameters at both sites, which suggests coccolithophores have kept pace with diatoms over the past 30 ka. The good correspondence also argues against preservational biases between biomarkers. The correlation is particularly strong at site ME0005A-15MC/17JC in the Eastern Pacific Warm Pool. Nutrients are much lower at this site and diatoms are consequently less abundant relative to other primary producers. It is therefore possible that haptophyte algae, rather than diatoms, are the dominant producer of brassicasterol at this site.

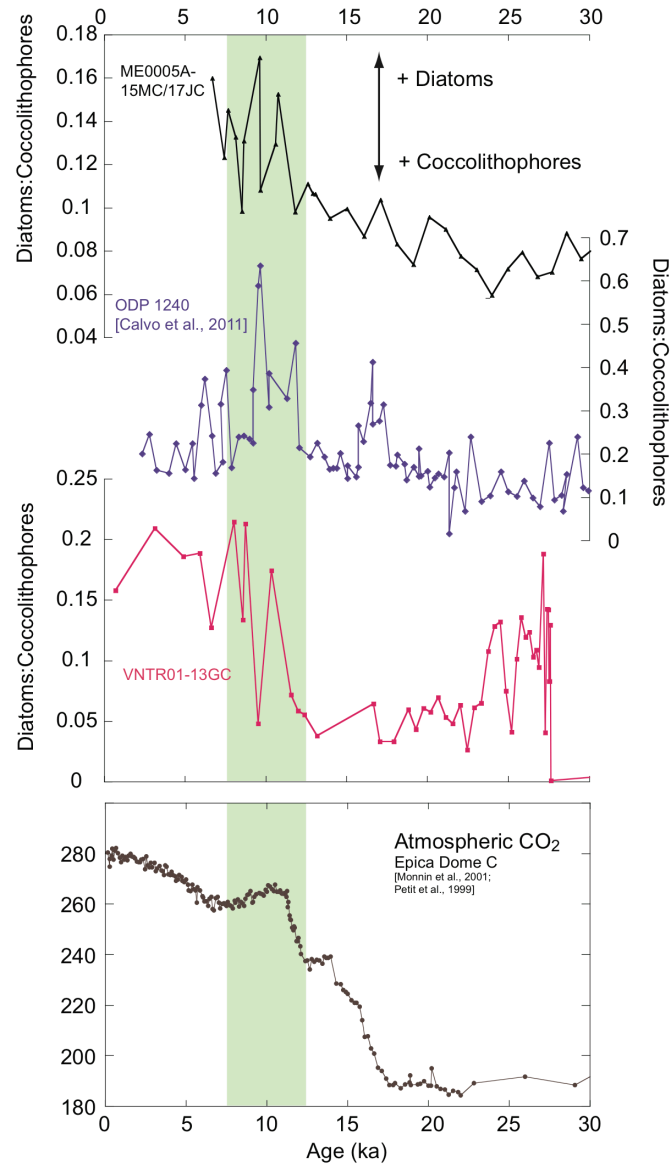


Figure 7. (top panel) The ratio of brassicasterol to C₃₇-alkenone abundances at three EEP open-ocean sites (black: ME0005A-15MC/17JC; purple: ODP 1240 [Calvo *et al.*, 2011]; pink: VNTR01-13GC) plotted with (bottom panel) the record of atmospheric CO₂ over the past 30 kyrs from Epica Dome C in Antarctica [Monnin *et al.*, 2001; Petit *et al.*, 1999]. Higher values indicate shifts towards increasingly diatom dominated production regimes. Diatoms compose a greater proportion of the total production at all sites

between 8 and 12 ka, though the increase is modest at sites ME0005A-15MC/17JC and VNTR01-13GC.

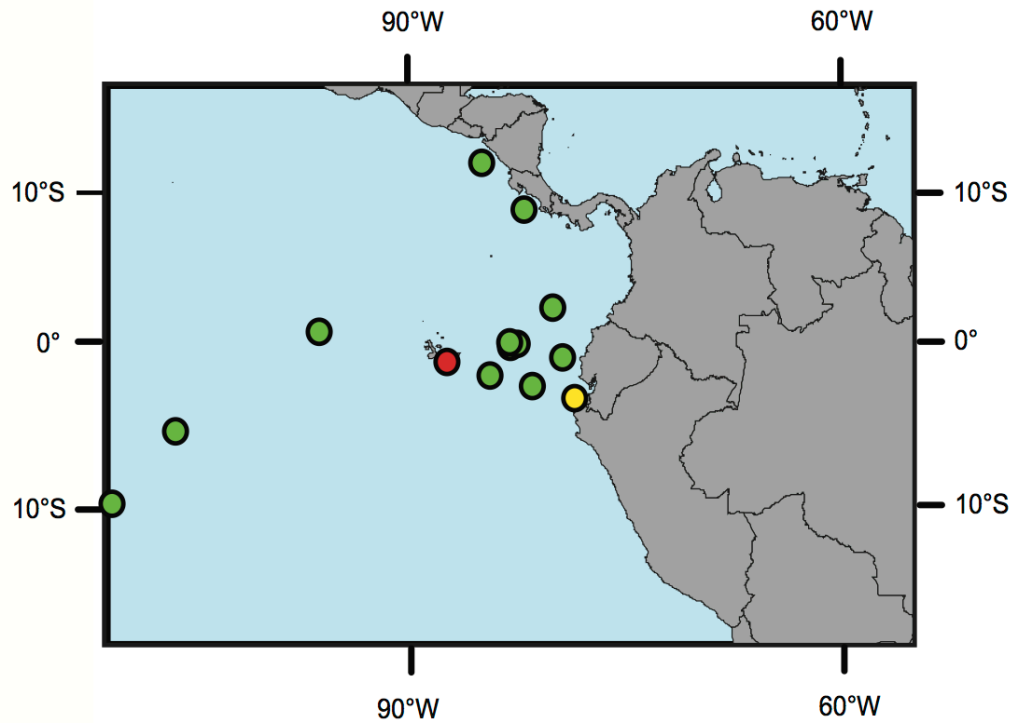


Figure 8. Same as Figure 1 but locations for inorganic proxy paleoproductivity records, including % calcite, % opal, and benthic foraminiferal abundances are removed. These inorganic proxy indicators likely reflect preservation rather than production [*Pichevin et al.*, 2009; *Dubois et al.*, 2010]. Red symbols indicate a decrease in productivity in the LGM relative to Holocene conditions. Green symbols indicate an increase in productivity in the LGM relative to Holocene conditions. Yellow symbols indicate no significant difference.

APPENDIX A: Organic Geochemical Proxy Data

CDH 23:

Age (Fairbanks Calibration)	Age (Marine13, Clam)	Depth (cm)	U_{37}^K	Temp. (°C, Prahl)	Temp. (°C, Muller)	C_{37}^{total} (nmol/g)
1273	932	28	0.814	22.78	23.32	2.19
1273	932	28	0.815	22.82	23.36	2.35
1326	983	32	0.816	22.86	23.41	2.82
1326	983	32	0.822	23.02	23.56	6.74
1379	1034	36	0.817	22.87	23.42	3.57
1379	1034	36	0.818	22.92	23.46	5.77
1431	1085	40	0.825	23.11	23.66	3.07
1483	1136	44	0.822	23.02	23.56	2.60
1483	1136	44	0.828	23.19	23.74	6.83
1534	1186	48	0.818	22.92	23.46	2.81
1585	1236	52	0.813	22.77	23.31	2.80
1585	1236	52	0.820	22.98	23.53	7.62
1636	1286	56	0.809	22.66	23.19	2.87
1687	1335	60	0.808	22.61	23.14	2.02
1687	1335	60	0.814	22.79	23.33	6.23
1737	1384	64	0.809	22.63	23.17	2.29
1786	1433	68	0.821	22.99	23.53	7.71
1786	1433	68	0.819	22.94	23.48	7.71
1786	1433	68	0.811	22.71	23.25	2.84
1786	1433	68	0.814	22.78	23.32	5.00
1836	1481	72	0.815	22.82	23.36	3.46
1836	1481	72	0.811	22.71	23.25	2.69
1836	1481	72	0.814	22.79	23.33	2.31
1885	1530	76	0.807	22.59	23.12	2.64
1885	1530	76	0.818	22.90	23.44	6.69
1934	1578	80	0.805	22.53	23.06	4.07
1934	1578	80	0.812	22.73	23.27	6.31
1983	1625	84	0.804	22.50	23.03	4.11
1983	1625	84	0.808	22.63	23.16	5.21
2031	1672	88	0.802	22.45	22.98	4.02
2079	1719	92	0.800	22.37	22.90	3.04
2079	1719	92	0.806	22.55	23.08	7.33
2126	1766	96	0.800	22.37	22.90	3.12
2174	1813	100	0.804	22.50	23.03	5.35
2174	1813	100	0.801	22.42	22.95	3.59
2221	1859	104	0.802	22.44	22.96	3.37
2267	1905	108	0.803	22.48	23.01	3.67
2314	1950	112	0.802	22.44	22.97	3.68
2314	1950	112	0.808	22.62	23.15	4.43
2360	1996	116	0.803	22.48	23.01	3.63
2360	1996	116	0.809	22.66	23.19	4.36
2383	2018	118	0.810	22.66	23.20	7.08
2406	2041	120	0.803	22.47	23.00	5.62

2406	2041	120	0.804	22.51	23.04	4.63
2452	2086	124	0.807	22.58	23.11	4.45
2497	2131	128	0.808	22.61	23.14	5.41
2542	2175	132	0.805	22.52	23.05	4.78
2587	2219	136	0.802	22.45	22.98	6.11
2631	2263	140	0.805	22.54	23.08	5.37
2676	2307	144	0.808	22.62	23.15	4.03
2720	2350	148	0.811	22.70	23.24	5.97
2763	2393	152	0.810	22.67	23.20	4.46
2763	2393	152	0.813	22.77	23.31	4.38
2807	2436	156	0.813	22.75	23.29	3.89
2850	2479	160	0.813	22.77	23.31	3.59
2893	2521	164	0.814	22.81	23.35	3.93
2936	2564	168	0.813	22.78	23.32	4.40
2979	2606	172	0.812	22.74	23.28	3.51
3021	2648	176	0.812	22.74	23.28	3.93
3063	2689	180	0.815	22.82	23.36	3.83
3063	2689	180	0.822	23.03	23.57	6.03
3105	2731	184	0.820	22.97	23.51	6.48
3105	2731	184	0.816	22.85	23.39	6.60
3105	2731	184	0.821	23.01	23.56	5.52
3105	2731	184	0.823	23.05	23.60	5.61
3147	2772	188	0.817	22.88	23.42	7.26
3188	2813	192	0.818	22.92	23.46	6.56
3229	2854	196	0.820	22.97	23.52	6.38
3270	2894	200	0.820	22.96	23.51	7.18
3291	2915	202	0.817	22.87	23.42	6.87
3291	2915	202	0.817	22.89	23.43	7.05
3311	2935	204	0.824	23.10	23.65	4.31
3352	2975	208	0.823	23.05	23.60	6.97
3392	3015	212	0.820	22.97	23.51	6.26
3432	3055	216	0.818	22.90	23.44	4.97
3472	3095	220	0.819	22.94	23.49	6.39
3472	3095	220	0.819	22.93	23.47	6.31
3482	3104	221	0.818	22.90	23.44	6.27
3512	3134	224	0.819	22.93	23.47	6.79
3552	3173	228	0.817	22.88	23.42	6.52
3591	3212	232	0.825	23.12	23.67	6.07
3630	3251	236	0.822	23.03	23.58	6.29
3669	3290	240	0.821	23.00	23.55	5.95
3708	3329	244	0.822	23.02	23.57	5.23
3747	3367	248	0.823	23.05	23.60	
3756	3377	249	0.821	22.99	23.53	7.35
3785	3405	252	0.821	23.01	23.56	6.68
3824	3444	256	0.824	23.07	23.62	6.24
3900	3519	264	0.821	22.99	23.54	6.40
3900	3519	264	0.814	22.80	23.34	4.36
3900	3519	264	0.816	22.85	23.39	5.00
3938	3557	268	0.815	22.83	23.37	4.61

3975	3594	272	0.807	22.60	23.13	4.20
4013	3632	276	0.810	22.66	23.20	5.06
4050	3669	280	0.817	22.88	23.42	5.00
4087	3706	284	0.815	22.82	23.36	4.52
4124	3743	288	0.814	22.79	23.33	4.17
4161	3780	292	0.809	22.66	23.19	3.47
4198	3816	296	0.812	22.75	23.29	3.91
4198	3816	296	0.818	22.92	23.46	7.18
4235	3853	300	0.817	22.87	23.41	4.32
4235	3853	300	0.815	22.82	23.36	4.65
4271	3889	304	0.814	22.80	23.34	4.16
4308	3925	308	0.816	22.84	23.38	4.27
4344	3961	312	0.815	22.81	23.35	4.88
4380	3997	316	0.815	22.82	23.36	4.05
4416	4033	320	0.811	22.71	23.25	4.45
4416	4033	320	0.812	22.73	23.26	7.17
4452	4069	324	0.814	22.79	23.33	4.14
4461	4078	325	0.805	22.53	23.06	7.20
4487	4104	328	0.807	22.58	23.11	5.40
4523	4140	332	0.808	22.62	23.16	4.48
4541	4158	334	0.814	22.78	23.32	5.22
4541	4158	334	0.807	22.58	23.12	4.11
4576	4193	338	0.815	22.82	23.36	5.43
4611	4228	342	0.811	22.71	23.24	5.70
4647	4263	346	0.808	22.60	23.14	5.35
4682	4298	350	0.813	22.75	23.29	5.68
4682	4298	350	0.805	22.54	23.07	6.16
4682	4298	350	0.805	22.54	23.07	5.68
4708	4324	353	0.805	22.53	23.06	6.87
4717	4333	354	0.814	22.80	23.34	5.87
4751	4368	358	0.813	22.76	23.30	5.86
4786	4402	362	0.813	22.75	23.29	5.78
4786	4402	362	0.812	22.72	23.26	5.85
4821	4437	366	0.813	22.76	23.30	5.56
4855	4471	370	0.811	22.71	23.24	6.24
4890	4506	374	0.814	22.79	23.33	5.81
4924	4540	378	0.813	22.77	23.31	4.47
4958	4574	382	0.815	22.81	23.35	5.43
4993	4608	386	0.816	22.86	23.41	5.63
5027	4642	390	0.815	22.82	23.36	5.16
5044	4659	392	0.814	22.80	23.34	4.99
5044	4659	392	0.811	22.70	23.24	4.97
5052	4668	393	0.815	22.81	23.35	7.50
5078	4693	396	0.817	22.89	23.43	5.03
5112	4727	400	0.822	23.02	23.57	4.77
5145	4761	404	0.820	22.97	23.52	5.28
5179	4794	408	0.821	22.99	23.53	5.08
5213	4828	412	0.819	22.95	23.50	5.27
5238	4853	415	0.816	22.86	23.40	4.85

5246	4861	416	0.819	22.95	23.50	4.42
5280	4895	420	0.819	22.94	23.48	4.46
5288	4903	421	0.821	23.00	23.54	6.15
5288	4903	421	0.819	22.94	23.49	4.56
5313	4928	424	0.824	23.08	23.63	6.45
5313	4928	424	0.823	23.06	23.61	6.85
5346	4961	428	0.825	23.10	23.65	5.86
5380	4994	432	0.824	23.09	23.64	5.62
5413	5027	436	0.825	23.13	23.68	6.04
5446	5061	440	0.827	23.18	23.73	5.61
5463	5077	442	0.823	23.05	23.60	7.14
5479	5094	444	0.821	23.00	23.55	7.57
5479	5094	444	0.829	23.22	23.78	5.65
5512	5126	448	0.829	23.23	23.78	5.60
5545	5159	452	0.827	23.18	23.73	6.36
5578	5192	456	0.829	23.23	23.78	5.25
5578	5192	456	0.826	23.16	23.71	5.17
5611	5225	460	0.829	23.24	23.80	6.69
5644	5258	464	0.833	23.34	23.90	5.84
5676	5290	468	0.830	23.25	23.80	6.85
5709	5323	472	0.831	23.29	23.85	5.46
5742	5355	476	0.827	23.17	23.72	6.21
5774	5388	480	0.829	23.23	23.78	6.16
5807	5420	484	0.831	23.30	23.85	5.58
5840	5453	488	0.826	23.15	23.70	5.85
5872	5485	492	0.833	23.35	23.91	6.03
5872	5485	492	0.831	23.29	23.84	6.03
5904	5517	496	0.831	23.28	23.84	5.10
5937	5550	500	0.836	23.43	23.99	5.51
5969	5582	504	0.833	23.37	23.92	5.21
6002	5614	508	0.832	23.31	23.87	6.10
6034	5646	512	0.832	23.32	23.87	6.39
6066	5678	516	0.835	23.41	23.97	5.94
6098	5710	520	0.833	23.36	23.92	6.30
6131	5742	524	0.830	23.26	23.81	5.50
6163	5774	528	0.833	23.34	23.90	5.87
6195	5806	532	0.833	23.35	23.91	5.40
6211	5822	534	0.831	23.29	23.85	5.31
6211	5822	534	0.831	23.29	23.84	5.28
6243	5854	538	0.837	23.47	24.03	4.80
6275	5886	542	0.834	23.38	23.93	5.62
6307	5918	546	0.834	23.37	23.93	5.08
6339	5950	550	0.832	23.32	23.88	4.37
6372	5982	554	0.827	23.18	23.73	5.05
6404	6013	558	0.832	23.32	23.88	4.78
6436	6045	562	0.831	23.29	23.85	5.85
6468	6077	566	0.832	23.31	23.87	5.60
6500	6109	570	0.832	23.33	23.89	5.65
6508	6117	571	0.827	23.18	23.73	8.28

6532	6140	574	0.832	23.33	23.89	5.31
6564	6172	578	0.826	23.16	23.71	4.02
6564	6172	578	0.830	23.25	23.81	5.86
6595	6204	582	0.829	23.23	23.78	5.22
6627	6235	586	0.829	23.22	23.77	5.76
6659	6267	590	0.827	23.18	23.73	4.71
6691	6299	594	0.831	23.30	23.85	7.28
6691	6299	594	0.828	23.22	23.77	7.39
6723	6330	598	0.828	23.20	23.75	6.93
6755	6362	602	0.828	23.22	23.77	7.30
6763	6370	603	0.822	23.02	23.57	8.89
6763	6370	603	0.821	22.99	23.54	8.48
6787	6394	606	0.818	22.90	23.45	6.88
6819	6425	610	0.823	23.07	23.62	6.96
6851	6457	614	0.823	23.07	23.62	7.63
6883	6488	618	0.824	23.10	23.65	6.55
6883	6488	618	0.823	23.07	23.62	7.06
6915	6520	622	0.823	23.07	23.62	7.50
6947	6552	626	0.821	23.01	23.55	7.12
6979	6583	630	0.823	23.06	23.61	7.85
7011	6615	634	0.822	23.03	23.58	6.86
7043	6646	638	0.820	22.97	23.51	6.28
7074	6678	642	0.820	22.98	23.53	6.03
7106	6709	646	0.819	22.94	23.49	5.31
7106	6709	646	0.820	22.96	23.50	5.52
7138	6741	650	0.818	22.90	23.44	2.88
7170	6773	654	0.819	22.93	23.48	3.96
7202	6804	658	0.819	22.95	23.49	5.13
7234	6836	662	0.819	22.95	23.50	6.52
7266	6867	666	0.819	22.95	23.50	5.59
7298	6899	670	0.822	23.04	23.59	5.68
7330	6930	674	0.824	23.08	23.62	4.56
7362	6962	678	0.824	23.10	23.65	5.47
7394	6994	682	0.820	22.97	23.51	3.71
7426	7025	686	0.821	23.01	23.55	2.77
7442	7041	688	0.820	22.98	23.53	5.13
7475	7072	692	0.811	22.71	23.24	6.22
7475	7072	692	0.810	22.67	23.20	6.49
7507	7104	696	0.815	22.81	23.35	5.86
7539	7136	700	0.816	22.85	23.40	5.19
7555	7152	702	0.805	22.52	23.05	6.20
7571	7167	704	0.813	22.76	23.29	5.23
7603	7199	708	0.808	22.62	23.15	5.63
7635	7231	712	0.807	22.58	23.12	6.81
7667	7262	716	0.807	22.58	23.12	6.41
7700	7294	720	0.804	22.50	23.03	5.79
7732	7326	724	0.803	22.46	22.99	6.17
7764	7357	728	0.805	22.52	23.05	3.64
7796	7389	732	0.801	22.41	22.94	6.22

7829	7421	736	0.800	22.37	22.90	3.65
7829	7421	736	0.800	22.40	22.92	3.76
7861	7452	740	0.801	22.41	22.93	4.56
7893	7484	744	0.797	22.28	22.81	4.23
7925	7516	748	0.797	22.28	22.80	3.99
7958	7548	752	0.801	22.41	22.94	3.94
7990	7579	756	0.803	22.48	23.01	4.00
8023	7611	760	0.799	22.36	22.88	4.86
8023	7611	760	0.796	22.25	22.78	4.88
8055	7643	764	0.802	22.44	22.97	3.74
8087	7675	768	0.802	22.44	22.96	3.50
8120	7707	772	0.794	22.20	22.72	3.54
8152	7739	776	0.794	22.21	22.73	3.46
8217	7802	784	0.790	22.08	22.59	4.64
8274	7858	791	0.795	22.25	22.77	6.24
8283	7866	792	0.789	22.07	22.59	4.85
8315	7898	796	0.790	22.10	22.61	4.58
8348	7930	800	0.794	22.21	22.73	4.61
8380	7962	804	0.795	22.23	22.76	4.71
8413	7994	808	0.796	22.28	22.80	4.44
8446	8026	812	0.791	22.11	22.63	4.49
8479	8058	816	0.797	22.29	22.81	4.69
8511	8090	820	0.801	22.40	22.92	4.26
8511	8090	820	0.801	22.41	22.93	4.35
8536	8114	823	0.791	22.12	22.64	3.54
8544	8122	824	0.794	22.22	22.74	5.06
8552	8130	825	0.795	22.24	22.76	5.15
8577	8154	828	0.794	22.20	22.72	4.78
8610	8186	832	0.790	22.10	22.61	4.81
8642	8218	836	0.785	21.93	22.44	5.63
8675	8250	840	0.785	21.95	22.46	3.96
8675	8250	840	0.784	21.92	22.43	3.88
8708	8283	844	0.786	21.98	22.49	4.26
8741	8315	848	0.782	21.85	22.36	4.21
8774	8347	852	0.778	21.74	22.25	3.92
8807	8379	856	0.779	21.77	22.28	4.26
8840	8411	860	0.780	21.78	22.29	3.70
8873	8443	864	0.792	22.14	22.66	3.80
8906	8476	868	0.781	21.82	22.33	3.63
8939	8508	872	0.777	21.71	22.22	4.14
8947	8516	873	0.790	22.10	22.62	5.02
8972	8540	876	0.780	21.80	22.31	3.66
9005	8572	880	0.781	21.83	22.34	3.75
9038	8605	884	0.781	21.82	22.33	3.85
9038	8605	884	0.781	21.83	22.34	3.74
9071	8637	888	0.782	21.87	22.38	3.12
9104	8669	892	0.779	21.76	22.26	4.17
9137	8702	896	0.779	21.77	22.27	3.97
9170	8734	900	0.780	21.80	22.31	3.54

9204	8766	904	0.785	21.94	22.45	3.33
9237	8799	908	0.784	21.91	22.43	3.45
9270	8831	912	0.778	21.72	22.23	3.91
9303	8864	916	0.779	21.77	22.28	3.71
9303	8864	916	0.778	21.74	22.25	3.74
9336	8896	920	0.783	21.89	22.40	3.11
9370	8928	924	0.782	21.84	22.35	2.82
9403	8961	928	0.779	21.76	22.27	4.42
9403	8961	928	0.777	21.71	22.21	4.29
9436	8993	932	0.784	21.90	22.41	3.35
9469	9026	936	0.782	21.87	22.38	3.81
9503	9058	940	0.780	21.80	22.31	3.37
9536	9091	944	0.778	21.75	22.26	4.43
9569	9123	948	0.785	21.94	22.46	3.74
9603	9156	952	0.785	21.94	22.46	3.89
9636	9188	956	0.778	21.74	22.25	4.38
9669	9221	960	0.777	21.71	22.21	3.47
9703	9253	964	0.778	21.74	22.25	4.45
9703	9253	964	0.776	21.68	22.19	4.29
9736	9286	968	0.774	21.62	22.12	3.99
9769	9318	972	0.777	21.69	22.20	4.00
9803	9351	976	0.777	21.70	22.20	3.44
9836	9383	980	0.774	21.62	22.13	3.94
9869	9416	984	0.778	21.74	22.25	3.85
9903	9448	988	0.776	21.68	22.19	2.94
9936	9481	992	0.772	21.57	22.07	3.84
9953	9497	994	0.767	21.40	21.90	2.87
9953	9497	994	0.765	21.35	21.85	2.84
9986	9530	998	0.773	21.59	22.09	3.50
10020	9562	1002	0.773	21.59	22.09	3.76
10020	9562	1002	0.772	21.57	22.07	3.80
10086	9627	1010	0.774	21.61	22.12	3.34
10120	9660	1014	0.769	21.48	21.98	3.74
10153	9693	1018	0.765	21.37	21.86	3.32
10220	9758	1026	0.771	21.53	22.03	3.11
10253	9790	1030	0.771	21.52	22.02	3.20
10286	9823	1034	0.778	21.74	22.24	3.51
10320	9855	1038	0.780	21.79	22.29	2.99
10320	9855	1038	0.778	21.74	22.25	3.09
10353	9888	1042	0.788	22.03	22.55	3.05
10386	9920	1046	0.786	21.97	22.49	2.89
10386	9920	1046	0.789	22.05	22.57	3.31
10419	9953	1050	0.789	22.05	22.57	2.92
10419	9953	1050	0.785	21.93	22.44	3.10
10453	9985	1054	0.783	21.87	22.39	3.24
10453	9985	1054	0.780	21.79	22.30	3.40
10486	10018	1058	0.786	21.97	22.48	3.17
10519	10050	1062	0.788	22.02	22.53	2.79
10519	10050	1062	0.786	21.98	22.50	2.99

10585	10115	1070	0.779	21.76	22.27	3.19
10585	10115	1070	0.785	21.95	22.46	3.01
10619	10147	1074	0.779	21.78	22.29	2.95
10652	10180	1078	0.783	21.88	22.39	3.23
10685	10212	1082	0.785	21.93	22.45	2.98
10718	10244	1086	0.781	21.82	22.33	2.82
10751	10277	1090	0.784	21.91	22.42	2.94
10784	10309	1094	0.787	22.00	22.51	2.73
10817	10341	1098	0.786	21.96	22.48	3.07
10850	10374	1102	0.788	22.02	22.54	2.61
10883	10406	1106	0.788	22.02	22.53	1.95
10916	10438	1110	0.784	21.90	22.42	2.69
10948	10470	1114	0.788	22.04	22.56	2.62
10981	10502	1118	0.786	21.98	22.49	2.29
11014	10534	1122	0.782	21.85	22.36	2.12
11047	10566	1126	0.789	22.07	22.59	1.80
11079	10599	1130	0.782	21.86	22.37	2.18
11144	10662	1138	0.784	21.92	22.43	2.18
11209	10726	1146	0.780	21.79	22.30	2.38
11218	10734	1147	0.781	21.84	22.35	2.27
11250	10766	1151	0.783	21.89	22.40	2.29
11282	10798	1155	0.780	21.79	22.30	2.25
11347	10861	1163	0.773	21.59	22.09	2.98
11347	10861	1163	0.771	21.52	22.02	2.94
11379	10893	1167	0.773	21.60	22.10	2.82
11411	10925	1171	0.769	21.48	21.97	2.39
11411	10925	1171	0.771	21.54	22.04	2.45
11507	11019	1183	0.770	21.50	22.00	3.11
11539	11051	1187	0.760	21.21	21.70	2.67
11570	11082	1191	0.763	21.30	21.80	1.76
11602	11113	1195	0.763	21.29	21.78	2.43
11634	11145	1199	0.763	21.31	21.80	2.97
11665	11176	1203	0.763	21.30	21.80	3.07
11665	11176	1203	0.765	21.36	21.85	3.05
11665	11176	1203	0.765	21.36	21.86	3.00
11697	11207	1207	0.759	21.17	21.66	3.50
11759	11269	1215	0.761	21.24	21.74	3.50
11791	11300	1219	0.766	21.39	21.89	3.09
11791	11300	1219	0.763	21.29	21.78	3.12
11822	11331	1223	0.756	21.08	21.57	2.92
11853	11362	1227	0.755	21.05	21.54	2.98
11884	11392	1231	0.750	20.91	21.40	3.21
11884	11392	1231	0.752	20.96	21.45	3.48
11914	11423	1235	0.757	21.13	21.61	3.09
11914	11423	1235	0.761	21.23	21.72	3.06
11914	11423	1235	0.757	21.11	21.60	3.06
11945	11453	1239	0.755	21.07	21.55	2.92
11945	11453	1239	0.763	21.29	21.78	2.82
11976	11484	1243	0.752	20.98	21.46	2.88

12006	11514	1247	0.755	21.07	21.56	1.95
12006	11514	1247	0.759	21.18	21.67	2.80
12037	11545	1251	0.759	21.18	21.67	2.31
12037	11545	1251	0.750	20.91	21.39	3.57
12060	11567	1254	0.752	20.98	21.47	3.58
12097	11605	1259	0.755	21.04	21.53	2.68
12127	11635	1263	0.761	21.25	21.74	2.73
12127	11635	1263	0.756	21.09	21.58	2.69
12127	11635	1263	0.754	21.02	21.50	2.73
12157	11665	1267	0.754	21.04	21.53	2.53
12210	11717	1274	0.752	20.96	21.45	3.84
12217	11725	1275	0.747	20.83	21.31	2.76
12276	11784	1283	0.747	20.83	21.31	2.96
12276	11784	1283	0.751	20.95	21.44	2.95
12305	11814	1287	0.750	20.90	21.38	3.07
12335	11843	1291	0.753	20.99	21.47	3.95
12364	11872	1295	0.749	20.89	21.38	
12393	11901	1299	0.753	21.01	21.50	3.07
12429	11938	1304	0.755	21.06	21.55	3.36
12486	11996	1312	0.753	21.01	21.49	3.40
12514	12024	1316	0.754	21.03	21.51	3.17
12543	12053	1320	0.752	20.98	21.46	2.41
12571	12081	1324	0.755	21.05	21.54	3.36
12626	12138	1332	0.736	20.51	20.98	2.91
12654	12166	1336	0.738	20.56	21.03	2.61
12682	12194	1340	0.729	20.30	20.76	3.47
12709	12222	1344	0.731	20.35	20.82	3.05
12736	12250	1348	0.738	20.56	21.03	2.93
12763	12277	1352	0.748	20.84	21.32	2.28
12790	12305	1356	0.745	20.78	21.25	2.04
12816	12332	1360	0.752	20.96	21.44	2.23
12843	12359	1364	0.754	21.04	21.52	2.02
12869	12386	1368	0.752	20.97	21.46	2.33
12895	12413	1372	0.746	20.79	21.27	2.24
12921	12440	1376	0.744	20.74	21.21	3.02
12947	12466	1380	0.750	20.90	21.38	2.66
12972	12493	1384	0.739	20.58	21.05	
12998	12519	1388	0.739	20.57	21.05	3.15
13023	12545	1392	0.742	20.68	21.16	2.56
13023	12545	1392	0.738	20.56	21.03	2.58
13048	12571	1396	0.744	20.73	21.20	2.47
13072	12597	1400	0.746	20.79	21.27	2.26
13097	12623	1404	0.746	20.78	21.26	2.97
13097	12623	1404	0.746	20.78	21.26	2.92
13121	12648	1408	0.749	20.88	21.36	2.90
13145	12673	1412	0.746	20.78	21.26	2.60
13145	12673	1412	0.747	20.83	21.31	2.51
13169	12698	1416	0.742	20.67	21.14	1.88
13193	12723	1420	0.747	20.81	21.29	2.05

13193	12723	1420	0.747	20.82	21.30	2.38
13216	12748	1424	0.747	20.83	21.31	1.66
13239	12773	1428	0.743	20.70	21.18	2.50
13262	12797	1432	0.748	20.86	21.34	2.48
13285	12821	1436	0.744	20.73	21.21	2.35
13308	12845	1440	0.750	20.91	21.40	
13324	12863	1443	0.744	20.72	21.20	2.12
13352	12893	1448	0.751	20.94	21.42	
13374	12916	1452	0.744	20.73	21.20	1.83
13395	12939	1456	0.745	20.77	21.25	1.91
13417	12963	1460	0.746	20.81	21.29	2.14
13443	12991	1465	0.739	20.59	21.06	1.76
13463	13014	1469	0.742	20.69	21.16	2.00
13463	13014	1469	0.743	20.72	21.20	2.00
13484	13036	1473	0.730	20.33	20.79	1.71
13499	13053	1476	0.737	20.54	21.01	2.20
13524	13080	1481	0.741	20.66	21.13	1.83
13544	13102	1485	0.741	20.66	21.13	2.11
13563	13123	1489	0.744	20.74	21.22	1.98
13582	13144	1493	0.742	20.68	21.16	1.92
13601	13165	1497	0.733	20.40	20.87	2.15
13619	13186	1501	0.734	20.44	20.90	1.86
13637	13207	1505	0.732	20.39	20.85	1.99
13655	13227	1509	0.747	20.81	21.29	2.58
13655	13227	1509	0.745	20.77	21.25	2.15
13669	13242	1512	0.751	20.93	21.41	2.35
13686	13262	1516	0.743	20.69	21.17	1.79
13690	13267	1517	0.743	20.71	21.19	1.35
13690	13267	1517	0.742	20.69	21.17	1.27
13703	13282	1520	0.735	20.47	20.93	2.26
13720	13301	1524	0.743	20.72	21.20	1.99
13724	13306	1525	0.748	20.86	21.34	1.13
13740	13325	1529	0.749	20.88	21.36	2.32
13752	13339	1532	0.739	20.58	21.05	2.15
13772	13362	1537	0.750	20.90	21.39	1.89
13784	13376	1540	0.739	20.59	21.07	2.13
13784	13376	1540	0.737	20.52	20.99	2.17
13803	13399	1545	0.741	20.65	21.13	2.40
13814	13412	1548	0.746	20.79	21.27	1.90
13832	13434	1553	0.745	20.77	21.25	2.15
13846	13451	1557	0.740	20.62	21.10	1.77
13860	13468	1561	0.733	20.41	20.88	1.84
13870	13481	1564	0.742	20.68	21.15	1.97
13873	13485	1565	0.737	20.52	20.99	2.01
13873	13485	1565	0.738	20.55	21.03	1.94
13886	13501	1569	0.751	20.95	21.44	0.63
13896	13513	1572	0.747	20.81	21.29	1.96
13899	13517	1573	0.747	20.83	21.31	2.30
13911	13533	1577	0.756	21.08	21.57	1.81

13920	13545	1580	0.750	20.90	21.39	1.76
13935	13564	1585	0.750	20.90	21.38	
13946	13578	1589	0.752	20.97	21.45	1.98
13957	13593	1593	0.741	20.66	21.13	1.49
13965	13604	1596	0.732	20.38	20.85	2.03
13977	13621	1601	0.738	20.56	21.03	1.91
13980	13625	1602	0.737	20.52	20.99	1.75
13982	13628	1603	0.749	20.87	21.35	2.28
13985	13632	1604	0.736	20.51	20.98	2.18
13994	13645	1608	0.742	20.68	21.16	1.81
14003	13658	1612	0.727	20.23	20.69	2.31
14012	13671	1616	0.730	20.33	20.80	
14020	13683	1620	0.731	20.35	20.81	2.01
14035	13707	1628	0.730	20.33	20.79	2.07
14042	13719	1632	0.749	20.88	21.36	1.57
14045	13724	1634	0.731	20.34	20.81	2.34
14045	13724	1634	0.732	20.38	20.84	2.41
14048	13730	1636	0.748	20.85	21.33	1.68
14055	13740	1640	0.756	21.10	21.59	1.55
14060	13751	1644	0.751	20.95	21.43	1.86
14060	13751	1644	0.751	20.95	21.43	1.71
14066	13761	1648	0.754	21.02	21.51	1.98
14070	13770	1652	0.754	21.04	21.53	1.72
14079	13789	1660	0.770	21.49	21.99	1.63
14082	13797	1664	0.761	21.22	21.72	1.83
14085	13806	1668	0.760	21.21	21.70	1.82
14088	13813	1672	0.766	21.38	21.88	1.60
14090	13821	1676	0.760	21.19	21.68	1.88
14091	13828	1680	0.765	21.35	21.84	1.60
14093	13835	1684	0.769	21.46	21.96	1.50
14093	13841	1688	0.775	21.66	22.16	1.31
14094	13847	1692	0.776	21.68	22.19	1.35
14093	13852	1696	0.768	21.43	21.93	1.40
14092	13857	1700	0.770	21.50	22.00	1.63
14089	13866	1708	0.768	21.43	21.93	1.52
14087	13870	1712	0.771	21.52	22.02	1.64
14086	13872	1714	0.770	21.51	22.01	1.58

CDH 26:

Age (yrs, Fairbanks calibration)	Age (yrs, Marine 13, Clam)	Depth (cm)	U_{37}^k	SST (°C, Prah)	Temp (°C, Muller)	C_{37}^{total} (nmol/g)
1366	1732	16	0.808	22.62	23.15	3.69
1924	2175	32	0.794	22.20	22.72	2.79
2460	2605	48	0.786	21.96	22.47	3.06
2976	3024	64	0.792	22.14	22.66	3.71
3472	3430	80	0.791	22.11	22.63	2.97
3948	3826	96	0.794	22.20	22.72	2.71
4406	4210	112	0.797	22.29	22.82	3.28
4846	4583	128	0.787	22.01	22.53	3.47

5562	5199	155.5	0.805	22.53	23.06	4.52
5956	5544	171.5	0.809	22.65	23.18	4.40
6334	5878	187.5	0.810	22.68	23.22	4.58
6698	6203	203.5	0.811	22.71	23.24	5.25
7047	6518	219.5	0.807	22.60	23.14	6.09
7382	6825	235.5	0.805	22.52	23.06	4.18
7703	7122	251.5	0.802	22.44	22.96	5.21
8012	7410	267.5	0.790	22.10	22.62	4.67
8308	7690	283.5	0.786	21.98	22.49	4.28
8478	7852	293	0.828	23.19	23.75	2.82
8755	8118	309	0.789	22.07	22.59	2.04
8840	8200	314	0.776	21.69	22.19	5.37
8840	8200	314	0.776	21.67	22.17	4.93
8906	8265	318	0.781	21.84	22.35	5.56
8973	8329	322	0.779	21.75	22.26	4.91
9103	8456	330	0.779	21.75	22.26	5.13
9167	8519	334	0.774	21.62	22.12	4.50
9167	8519	334	0.774	21.61	22.11	4.61
9230	8581	338	0.773	21.58	22.09	5.09
9355	8705	346	0.773	21.58	22.08	4.89
9417	8766	350	0.770	21.51	22.01	6.45
9477	8826	354	0.774	21.62	22.13	4.00
9597	8946	362	0.775	21.64	22.14	4.06
9715	9064	370	0.772	21.55	22.05	3.73
9715	9064	370	0.771	21.52	22.03	3.81
9830	9180	378	0.766	21.37	21.86	3.12
9887	9237	382	0.767	21.41	21.91	3.34
9943	9294	386	0.765	21.36	21.86	1.98
10054	9407	394	0.764	21.33	21.82	3.40
10108	9463	398	0.771	21.54	22.04	3.09
10108	9463	398	0.773	21.59	22.10	2.92
10162	9518	402	0.770	21.50	22.00	2.68
10268	9627	410	0.778	21.75	22.26	3.08
10321	9681	414	0.783	21.88	22.39	3.14
10373	9735	418	0.778	21.72	22.23	3.30
10475	9841	426	0.779	21.77	22.28	3.27
10525	9893	430	0.788	22.04	22.56	3.08
10575	9945	434	0.784	21.91	22.42	3.10
10575	9945	434	0.780	21.81	22.32	3.14
10673	10048	442	0.784	21.91	22.42	2.85
10770	10149	450	0.790	22.09	22.60	2.95
10817	10200	454	0.783	21.88	22.39	3.26
10864	10249	458	0.783	21.88	22.39	2.31
10864	10249	458	0.785	21.95	22.46	2.83
10968	10360	467	0.771	21.53	22.03	2.49
11014	10408	471	0.776	21.68	22.18	2.77
11059	10457	475	0.774	21.60	22.11	2.61
11148	10552	483	0.771	21.54	22.04	2.90
11192	10599	487	0.778	21.74	22.24	2.91

11235	10646	491	0.773	21.58	22.09	2.93
11321	10738	499	0.777	21.70	22.20	2.81
11321	10738	499	0.774	21.63	22.13	2.95
11364	10784	503	0.766	21.38	21.88	2.37
11405	10829	507	0.772	21.56	22.07	3.04
11488	10919	515	0.769	21.46	21.96	2.77
11529	10964	519	0.766	21.38	21.88	2.84
11569	11008	523	0.767	21.41	21.90	2.80
11569	11008	523	0.766	21.38	21.87	2.55
11649	11095	531	0.759	21.17	21.66	2.92
11688	11138	535	0.762	21.27	21.77	2.87
11727	11181	539	0.757	21.10	21.59	3.22
11804	11266	547	0.763	21.30	21.80	3.44
11842	11308	551	0.755	21.06	21.54	2.54
11880	11349	555	0.757	21.11	21.60	3.00
11954	11431	563	0.749	20.89	21.38	4.23
11954	11431	563	0.748	20.85	21.33	4.22
11991	11472	567	0.760	21.20	21.69	3.50
12027	11513	571	0.749	20.90	21.38	4.31
12099	11593	579	0.741	20.65	21.12	3.32
12170	11671	587	0.736	20.49	20.96	4.65
12239	11749	595	0.738	20.55	21.02	4.30
12316	11835	604	0.753	20.99	21.47	4.72
12350	11873	608	0.752	20.96	21.44	4.55
12383	11911	612	0.741	20.65	21.12	4.68
12449	11985	620	0.735	20.48	20.95	4.11
12482	12022	624	0.735	20.48	20.95	3.91
12482	12022	624	0.735	20.48	20.95	3.89
12514	12059	628	0.725	20.17	20.63	4.39
12579	12131	636	0.734	20.46	20.92	4.55
12579	12131	636	0.734	20.44	20.91	4.64
12610	12167	640	0.744	20.74	21.22	4.70
12642	12203	644	0.750	20.90	21.38	4.79
12704	12273	652	0.725	20.17	20.63	3.97
12735	12308	656	0.727	20.22	20.68	3.78
12735	12308	656	0.725	20.19	20.65	3.94
12766	12343	660	0.723	20.13	20.58	4.01
12826	12412	668	0.730	20.33	20.80	3.88
12856	12446	672	0.724	20.14	20.60	3.86
12856	12446	672	0.724	20.14	20.60	3.81
12886	12480	676	0.720	20.02	20.48	3.81
12945	12546	684	0.729	20.30	20.77	3.14
12974	12580	688	0.733	20.42	20.89	3.32
12974	12580	688	0.730	20.34	20.80	3.23
13003	12613	692	0.733	20.41	20.88	3.12
13061	12678	700	0.726	20.20	20.66	2.96
13089	12710	704	0.731	20.36	20.82	2.74
13118	12742	708	0.736	20.49	20.96	2.68
13174	12806	716	0.730	20.33	20.79	2.52

13202	12838	720	0.736	20.49	20.96	2.60
13229	12869	724	0.732	20.37	20.84	2.60
13284	12931	732	0.721	20.06	20.52	2.35
13311	12962	736	0.719	19.99	20.44	2.75
13339	12993	740	0.714	19.86	20.31	2.64
13392	13053	748	0.734	20.43	20.90	2.61
13392	13053	748	0.733	20.41	20.88	2.63
13419	13084	752	0.724	20.15	20.61	2.34
13432	13099	754	0.727	20.23	20.70	2.59
13465	13136	759	0.729	20.29	20.75	2.59
13492	13166	763	0.734	20.44	20.90	2.67
13511	13188	766	0.730	20.34	20.80	2.03
13550	13232	772	0.735	20.47	20.94	2.12
13583	13268	777	0.727	20.24	20.70	2.68
13608	13297	781	0.726	20.21	20.67	2.75
13660	13354	789	0.732	20.38	20.84	3.02
13685	13383	793	0.730	20.33	20.80	2.77
13710	13411	797	0.721	20.07	20.52	2.66
13761	13467	805	0.727	20.25	20.71	3.00
13786	13495	809	0.732	20.37	20.83	2.82
13811	13523	813	0.727	20.24	20.70	2.90
13860	13578	821	0.733	20.40	20.87	2.81
13860	13578	821	0.731	20.37	20.83	2.84
13885	13606	825	0.731	20.35	20.82	2.80
13909	13633	829	0.720	20.02	20.48	2.96
13909	13633	829	0.729	20.30	20.77	2.90
13909	13633	829	0.724	20.15	20.61	2.52
13934	13660	833	0.724	20.15	20.61	3.12
13958	13687	837	0.731	20.35	20.82	2.97
13983	13714	841	0.745	20.76	21.24	2.49
14007	13740	845	0.724	20.15	20.61	3.17
14031	13767	849	0.723	20.11	20.57	2.40
14055	13794	853	0.734	20.43	20.90	2.48
14079	13820	857	0.724	20.16	20.62	2.45
14103	13846	861	0.737	20.52	20.99	2.76
14127	13872	865	0.738	20.56	21.03	2.70
14151	13898	869	0.757	21.11	21.60	2.52
14175	13924	873	0.737	20.53	21.00	2.02
14199	13950	877	0.736	20.49	20.96	2.51
14223	13976	881	0.724	20.15	20.61	2.95
14246	14002	885	0.743	20.71	21.19	1.53
14270	14027	889	0.739	20.60	21.07	3.05
14294	14053	893	0.755	21.07	21.56	2.45
14317	14078	897	0.751	20.94	21.43	2.63
14317	14078	897	0.756	21.09	21.58	2.69
14341	14103	901	0.752	20.96	21.45	2.17
14364	14128	905	0.754	21.02	21.50	1.56
14364	14128	905	0.754	21.03	21.51	1.57
14387	14153	909	0.743	20.70	21.17	2.45

14411	14178	913	0.740	20.62	21.09	2.76
14434	14203	917	0.743	20.70	21.18	2.64
14457	14228	921	0.747	20.81	21.29	2.74
14481	14253	925	0.753	20.99	21.47	2.25
14504	14277	929	0.750	20.91	21.39	2.63
14539	14314	935	0.742	20.68	21.15	2.98
14550	14326	937	0.737	20.53	21.00	2.96
14573	14351	941	0.741	20.65	21.12	2.41
14597	14375	945	0.743	20.72	21.19	2.49
14620	14399	949	0.730	20.31	20.78	2.66
14643	14423	953	0.712	19.80	20.25	2.09
14666	14448	957	0.731	20.36	20.82	2.22
14689	14472	961	0.724	20.16	20.62	1.96
14689	14472	961	0.721	20.06	20.52	1.92
14712	14496	965	0.728	20.27	20.74	2.54
14735	14520	969	0.717	19.95	20.41	2.42
14758	14543	973	0.726	20.22	20.68	2.24
14781	14567	977	0.716	19.92	20.38	2.83
14804	14591	981	0.719	20.01	20.47	2.63
14827	14615	985	0.724	20.15	20.61	2.98
14850	14638	989	0.716	19.91	20.36	1.92
14873	14662	993	0.714	19.86	20.31	2.94
14896	14685	997	0.711	19.77	20.22	3.30
14919	14709	1001	0.715	19.90	20.35	3.86
14942	14732	1005	0.723	20.13	20.59	2.58
14965	14756	1009	0.706	19.63	20.07	2.74
14988	14779	1013	0.703	19.54	19.98	2.56
15011	14802	1017	0.711	19.75	20.20	2.70
15011	14802	1017	0.710	19.74	20.18	2.74
15034	14826	1021	0.716	19.90	20.36	3.02
15057	14849	1025	0.704	19.55	19.99	2.80
15080	14872	1029	0.701	19.48	19.92	2.93
15103	14895	1033	0.702	19.49	19.93	3.05
15126	14918	1037	0.705	19.58	20.02	3.18
15149	14941	1041	0.698	19.37	19.81	3.50
15172	14964	1045	0.701	19.47	19.91	3.31
15195	14987	1049	0.717	19.95	20.41	3.11
15218	15010	1053	0.710	19.73	20.17	2.94
15241	15033	1057	0.715	19.88	20.33	3.02
15241	15033	1057	0.713	19.82	20.27	3.07
15264	15056	1061	0.718	19.98	20.43	2.46
15287	15079	1065	0.722	20.08	20.54	3.23
15310	15102	1069	0.710	19.74	20.19	3.04
15333	15125	1073	0.717	19.93	20.38	3.53
15356	15147	1077	0.710	19.74	20.19	3.26
15356	15147	1077	0.711	19.77	20.22	3.42
15356	15147	1077	0.720	20.04	20.49	3.66
15380	15170	1081	0.726	20.22	20.68	3.30
15403	15193	1085	0.722	20.09	20.55	3.51

15426	15216	1089	0.720	20.02	20.48	3.82
15426	15216	1089	0.719	20.01	20.46	3.94
15449	15239	1093	0.732	20.39	20.86	3.67
15472	15261	1097	0.725	20.16	20.62	3.45
15496	15284	1101	0.714	19.85	20.30	3.52
15519	15307	1105	0.713	19.84	20.29	3.29
15542	15329	1109	0.723	20.11	20.57	3.05
15566	15352	1113	0.717	19.94	20.39	3.66
15589	15375	1117	0.726	20.19	20.65	3.28
15613	15397	1121	0.726	20.20	20.66	3.14
15636	15420	1125	0.713	19.82	20.27	3.79
15659	15443	1129	0.726	20.22	20.68	3.62
15683	15465	1133	0.716	19.91	20.36	3.45
15707	15488	1137	0.719	20.01	20.46	3.47
15707	15488	1137	0.720	20.03	20.48	3.48
15730	15511	1141	0.711	19.77	20.22	3.53
15754	15533	1145	0.717	19.94	20.39	4.24
15777	15556	1149	0.725	20.18	20.64	3.06
15801	15579	1153	0.728	20.27	20.73	3.30
15825	15601	1157	0.738	20.55	21.02	3.21
15848	15624	1161	0.732	20.38	20.84	3.13
15872	15647	1165	0.728	20.27	20.73	3.15
15896	15669	1169	0.728	20.26	20.72	2.96
15920	15692	1173	0.735	20.48	20.95	3.29
15968	15738	1181	0.731	20.34	20.81	3.18
15992	15760	1185	0.742	20.67	21.15	2.36
16016	15783	1189	0.720	20.02	20.47	3.29
16040	15806	1193	0.719	19.99	20.45	3.09
16064	15829	1197	0.721	20.07	20.52	2.42
16124	15886	1207	0.734	20.43	20.90	2.12
16160	15920	1213	0.708	19.67	20.11	2.92
16179	15937	1216	0.708	19.68	20.12	2.91
16203	15960	1220	0.705	19.58	20.02	1.97
16227	15983	1224	0.703	19.54	19.98	3.38
16252	16006	1228	0.708	19.66	20.11	3.30
16276	16029	1232	0.716	19.90	20.35	3.01
16301	16052	1236	0.715	19.88	20.33	2.42
16325	16075	1240	0.730	20.34	20.80	2.71
16374	16121	1248	0.712	19.79	20.24	3.34
16374	16121	1248	0.711	19.75	20.20	3.26
16399	16144	1252	0.717	19.95	20.40	2.97
16424	16167	1256	0.716	19.93	20.38	2.87
16448	16190	1260	0.720	20.03	20.49	0.85
16448	16190	1260	0.718	19.98	20.44	0.85
16473	16213	1264	0.721	20.06	20.51	3.26
16498	16237	1268	0.734	20.43	20.90	3.20
16523	16260	1272	0.737	20.53	21.00	2.48
16548	16283	1276	0.727	20.23	20.69	2.45
16573	16307	1280	0.716	19.90	20.35	3.01

16598	16330	1284	0.734	20.43	20.90	2.90
16623	16353	1288	0.736	20.50	20.97	2.95
16673	16400	1296	0.731	20.36	20.83	2.19
16723	16447	1304	0.727	20.22	20.68	2.68
16723	16447	1304	0.728	20.25	20.71	2.77
16749	16471	1308	0.719	20.00	20.46	3.08
16774	16494	1312	0.720	20.03	20.48	3.53
16799	16518	1316	0.720	20.02	20.47	2.95
16799	16518	1316	0.722	20.08	20.54	2.76
16825	16542	1320	0.734	20.46	20.92	2.44
16850	16566	1324	0.732	20.38	20.85	2.61
16876	16589	1328	0.736	20.49	20.95	3.30
16901	16613	1332	0.735	20.46	20.93	3.13
16978	16685	1344	0.733	20.40	20.87	3.71
17004	16709	1348	0.710	19.74	20.18	3.83
17030	16733	1352	0.733	20.41	20.88	3.27
17055	16757	1356	0.724	20.14	20.60	4.02
17081	16782	1360	0.720	20.04	20.50	4.27
17107	16806	1364	0.722	20.08	20.53	3.91
17140	16836	1369	0.720	20.04	20.49	4.67
17140	16836	1369	0.725	20.17	20.63	4.57
17166	16861	1373	0.714	19.87	20.32	3.35
17296	16983	1393	0.723	20.12	20.58	3.34
17322	17008	1397	0.720	20.04	20.49	2.41
17349	17033	1401	0.706	19.63	20.07	4.31
17375	17057	1405	0.726	20.19	20.65	2.60
17454	17132	1417	0.710	19.73	20.17	2.90
17481	17157	1421	0.714	19.85	20.30	3.88
17507	17182	1425	0.715	19.87	20.32	3.99
17534	17207	1429	0.720	20.04	20.49	4.07
17560	17232	1433	0.718	19.97	20.42	2.51
17560	17232	1433	0.718	19.96	20.41	2.49
17587	17258	1437	0.713	19.82	20.27	2.34
17614	17283	1441	0.733	20.43	20.89	4.03
17614	17283	1441	0.734	20.45	20.91	4.54
17647	17315	1446	0.710	19.73	20.17	2.64
17674	17340	1450	0.725	20.17	20.63	2.31
17755	17417	1462	0.706	19.61	20.05	4.17
17782	17442	1466	0.730	20.34	20.80	4.12
17809	17468	1470	0.735	20.47	20.94	1.81
17836	17494	1474	0.722	20.08	20.53	4.64
17863	17520	1478	0.724	20.16	20.62	2.54
17890	17546	1482	0.737	20.52	20.99	2.78
17917	17572	1486	0.728	20.26	20.72	4.67
17971	17624	1494	0.737	20.53	21.00	4.45
17998	17650	1498	0.727	20.24	20.71	4.70
18080	17729	1510	0.715	19.89	20.34	6.31
18108	17755	1514	0.713	19.83	20.28	5.93
18135	17782	1518	0.727	20.24	20.70	3.26

18163	17809	1522	0.732	20.37	20.84	4.07
18176	17822	1524	0.730	20.33	20.79	4.93
18176	17822	1524	0.729	20.30	20.76	5.13
18204	17848	1528	0.735	20.47	20.94	2.59
18204	17848	1528	0.735	20.47	20.93	2.58
18231	17875	1532	0.731	20.34	20.81	5.26
18259	17902	1536	0.722	20.09	20.55	2.80
18286	17929	1540	0.725	20.19	20.65	1.78
18314	17956	1544	0.722	20.10	20.55	4.41
18397	18037	1556	0.711	19.75	20.20	5.26
18425	18064	1560	0.709	19.70	20.15	5.81
18453	18091	1564	0.726	20.20	20.66	7.48
18480	18119	1568	0.728	20.27	20.74	7.85
18508	18146	1572	0.714	19.85	20.30	6.19
18536	18174	1576	0.719	20.01	20.46	5.75
18564	18201	1580	0.724	20.15	20.60	5.96
18592	18229	1584	0.728	20.26	20.72	6.59
18620	18256	1588	0.727	20.24	20.70	4.89
18648	18284	1592	0.715	19.89	20.34	5.80
18648	18284	1592	0.715	19.88	20.33	5.68
18676	18312	1596	0.711	19.77	20.22	6.77
18704	18340	1600	0.704	19.56	20.00	7.81
18760	18395	1608	0.712	19.78	20.23	6.17
18788	18423	1612	0.715	19.87	20.32	6.13
18788	18423	1612	0.715	19.87	20.32	6.05
18816	18452	1616	0.713	19.83	20.28	6.03
18872	18508	1624	0.724	20.16	20.62	5.75
18900	18536	1628	0.716	19.90	20.35	5.80
18929	18565	1632	0.707	19.64	20.08	5.50
18957	18593	1636	0.719	20.01	20.46	6.16
18985	18621	1640	0.707	19.64	20.09	5.72
19013	18650	1644	0.705	19.59	20.03	6.22
19070	18707	1652	0.721	20.07	20.52	5.67
19098	18736	1656	0.718	19.98	20.44	5.89
19126	18765	1660	0.730	20.32	20.78	5.52
19183	18823	1668	0.708	19.69	20.13	6.26
19212	18852	1672	0.715	19.89	20.34	7.20
19254	18895	1678	0.736	20.49	20.96	6.93
19254	18895	1678	0.741	20.65	21.12	7.43
19283	18924	1682	0.745	20.75	21.23	6.58
19283	18924	1682	0.742	20.67	21.14	6.30
19311	18954	1686	0.742	20.68	21.16	7.59
19339	18983	1690	0.737	20.54	21.01	8.22
19396	19041	1698	0.741	20.65	21.13	8.05
19425	19071	1702	0.740	20.61	21.08	7.02
19453	19100	1706	0.736	20.49	20.96	7.61
19482	19130	1710	0.730	20.32	20.79	7.87
19482	19130	1710	0.731	20.36	20.83	8.23
19510	19160	1714	0.742	20.69	21.16	8.35

19539	19189	1718	0.732	20.39	20.85	7.05
19568	19219	1722	0.742	20.68	21.16	5.56
19596	19249	1726	0.742	20.69	21.17	5.57
19625	19279	1730	0.743	20.69	21.17	5.25
19653	19308	1734	0.742	20.67	21.14	5.25
19710	19368	1742	0.735	20.47	20.93	5.12
19739	19398	1746	0.745	20.77	21.24	4.49
19768	19429	1750	0.724	20.14	20.60	5.94
19839	19504	1760	0.710	19.75	20.19	6.22
19868	19534	1764	0.707	19.64	20.09	5.74
19897	19565	1768	0.726	20.21	20.67	6.65
19897	19565	1768	0.719	20.00	20.45	6.35
19925	19595	1772	0.712	19.81	20.26	6.68
19925	19595	1772	0.709	19.69	20.14	6.26
19954	19626	1776	0.722	20.09	20.55	6.48
19983	19656	1780	0.724	20.15	20.61	6.83
20011	19687	1784	0.721	20.06	20.51	6.24
20040	19717	1788	0.735	20.47	20.94	5.29
20098	19779	1796	0.725	20.18	20.64	5.74
20126	19809	1800	0.720	20.03	20.48	6.22
20155	19840	1804	0.719	20.00	20.46	5.79
20184	19871	1808	0.719	20.00	20.46	3.68
20212	19902	1812	0.714	19.86	20.31	6.53
20241	19933	1816	0.717	19.95	20.40	6.71
20270	19964	1820	0.718	19.97	20.43	6.16
20299	19995	1824	0.733	20.42	20.89	6.65
20349	20049	1831	0.733	20.42	20.89	2.00
20349	20049	1831	0.733	20.41	20.88	1.96
20406	20111	1839	0.722	20.08	20.54	5.52
20435	20143	1843	0.727	20.24	20.70	5.68
20464	20174	1847	0.737	20.53	21.00	6.04
20464	20174	1847	0.732	20.37	20.84	5.36
20521	20236	1855	0.735	20.46	20.93	4.68
20550	20268	1859	0.725	20.16	20.62	5.36
20579	20299	1863	0.732	20.38	20.84	5.14
20608	20331	1867	0.718	19.96	20.41	4.82
20636	20362	1871	0.717	19.95	20.41	5.34
20723	20457	1883	0.736	20.51	20.98	5.82
20752	20488	1887	0.725	20.18	20.64	6.85
20752	20488	1887	0.728	20.25	20.72	7.10
20780	20520	1891	0.738	20.55	21.02	6.51
20809	20551	1895	0.740	20.62	21.09	6.78
20838	20583	1899	0.736	20.50	20.97	4.75
20867	20615	1903	0.738	20.56	21.03	6.06
20895	20647	1907	0.735	20.46	20.93	6.18
20982	20742	1919	0.725	20.19	20.65	5.45
21039	20806	1927	0.728	20.27	20.73	5.66
21039	20806	1927	0.728	20.25	20.72	5.59
21068	20837	1931	0.725	20.17	20.63	5.36

21097	20869	1935	0.729	20.28	20.74	5.42
21125	20901	1939	0.716	19.92	20.37	6.51
21154	20933	1943	0.721	20.06	20.52	6.06
21183	20965	1947	0.717	19.95	20.40	5.58
21212	20997	1951	0.735	20.47	20.94	6.01
21240	21029	1955	0.731	20.36	20.83	4.74
21269	21061	1959	0.728	20.27	20.74	4.63
21298	21093	1963	0.730	20.33	20.79	7.04
21327	21125	1967	0.730	20.32	20.78	6.78
21356	21157	1971	0.735	20.48	20.95	5.67
21384	21189	1975	0.732	20.37	20.84	5.57
21413	21221	1979	0.731	20.35	20.82	4.87
21456	21269	1985	0.724	20.16	20.62	4.21
21456	21269	1985	0.716	19.91	20.37	3.50
21485	21301	1989	0.724	20.15	20.61	5.24
21514	21333	1993	0.718	19.98	20.43	5.52
21542	21365	1997	0.733	20.42	20.88	5.38
21571	21397	2001	0.731	20.34	20.81	5.52
21571	21397	2001	0.733	20.42	20.89	5.27
21600	21429	2005	0.731	20.35	20.81	5.03
21686	21525	2017	0.712	19.80	20.25	4.09
21715	21557	2021	0.728	20.27	20.74	5.86
21744	21589	2025	0.726	20.22	20.68	5.36
21773	21621	2029	0.724	20.16	20.62	5.22
21801	21653	2033	0.725	20.18	20.64	5.24
21830	21685	2037	0.730	20.34	20.80	5.19
21859	21717	2041	0.733	20.42	20.89	5.46
21888	21749	2045	0.729	20.30	20.77	3.08
21917	21781	2049	0.716	19.92	20.37	4.20
21938	21804	2052	0.725	20.16	20.62	5.23
21974	21844	2057	0.730	20.32	20.78	7.09
22003	21876	2061	0.726	20.21	20.67	4.87
22032	21908	2065	0.743	20.70	21.17	6.11
22032	21908	2065	0.740	20.62	21.09	5.88
22061	21940	2069	0.742	20.68	21.15	4.57
22089	21972	2073	0.732	20.39	20.85	5.39
22089	21972	2073	0.727	20.22	20.68	4.77
22118	22003	2077	0.736	20.49	20.96	5.24
22147	22035	2081	0.729	20.31	20.77	4.89
22176	22067	2085	0.739	20.59	21.06	5.60
22205	22099	2089	0.734	20.45	20.92	4.44
22234	22130	2093	0.736	20.50	20.97	3.91
22263	22162	2097	0.723	20.12	20.58	5.12
22292	22193	2101	0.729	20.28	20.75	2.81
22320	22225	2105	0.736	20.50	20.97	5.47
22349	22257	2109	0.744	20.74	21.22	4.55
22378	22288	2113	0.736	20.50	20.97	4.66
22378	22288	2113	0.735	20.46	20.93	4.64
22407	22319	2117	0.739	20.59	21.06	5.16

22436	22351	2121	0.742	20.67	21.15	5.12
22465	22382	2125	0.733	20.41	20.87	5.30
22494	22414	2129	0.736	20.50	20.97	5.79
22523	22445	2133	0.742	20.69	21.17	3.90
22545	22468	2136	0.733	20.40	20.86	2.83
22574	22499	2140	0.727	20.23	20.69	6.02
22574	22499	2140	0.727	20.24	20.71	5.97
22603	22530	2144	0.728	20.27	20.73	5.35
22632	22562	2148	0.721	20.06	20.52	5.48
22661	22593	2152	0.719	20.00	20.45	5.95
22690	22624	2156	0.731	20.35	20.82	5.35
22719	22654	2160	0.727	20.24	20.71	5.53
22749	22685	2164	0.726	20.22	20.68	5.05
22778	22716	2168	0.740	20.60	21.08	5.27
22807	22747	2172	0.744	20.74	21.22	4.33
22836	22777	2176	0.743	20.71	21.18	4.39
22836	22777	2176	0.743	20.72	21.20	4.21
22866	22808	2180	0.733	20.42	20.89	5.61
22895	22839	2184	0.728	20.26	20.72	5.49
22924	22869	2188	0.735	20.48	20.95	5.66
22953	22899	2192	0.739	20.58	21.05	5.07
22953	22899	2192	0.739	20.59	21.06	5.01
22983	22930	2196	0.741	20.65	21.12	4.33
23012	22960	2200	0.744	20.72	21.20	2.70
23042	22990	2204	0.737	20.53	21.00	5.43
23071	23020	2208	0.735	20.48	20.95	6.17
23101	23050	2212	0.743	20.70	21.18	5.78
23130	23080	2216	0.745	20.77	21.25	6.14
23160	23110	2220	0.746	20.79	21.26	5.70
23189	23140	2224	0.744	20.73	21.21	4.94
23219	23169	2228	0.746	20.79	21.27	3.35
23249	23199	2232	0.742	20.68	21.16	5.32
23278	23229	2236	0.742	20.68	21.16	4.96
23308	23258	2240	0.738	20.55	21.02	5.71
23308	23258	2240	0.737	20.54	21.01	5.62
23338	23287	2244	0.757	21.13	21.62	4.49
23368	23316	2248	0.752	20.96	21.44	5.08
23398	23345	2252	0.745	20.76	21.24	5.09
23428	23374	2256	0.753	20.99	21.48	4.08
23458	23403	2260	0.749	20.87	21.35	4.45
23488	23432	2264	0.747	20.81	21.29	4.46
23518	23461	2268	0.743	20.70	21.17	4.34
23548	23489	2272	0.745	20.76	21.23	4.98
23578	23518	2276	0.746	20.78	21.26	5.15
23609	23546	2280	0.754	21.02	21.51	4.76
23639	23575	2284	0.760	21.21	21.70	4.36
23669	23603	2288	0.764	21.34	21.83	3.16
23685	23617	2290	0.759	21.19	21.68	2.46
23685	23617	2290	0.759	21.17	21.66	2.55

23715	23645	2294	0.758	21.13	21.62	4.21
23746	23672	2298	0.761	21.25	21.74	3.91
23746	23672	2298	0.760	21.22	21.71	4.02
23776	23700	2302	0.760	21.21	21.70	4.02
23807	23728	2306	0.747	20.81	21.29	4.27
23838	23755	2310	0.740	20.62	21.09	4.47
23869	23782	2314	0.745	20.78	21.26	4.32
23869	23782	2314	0.746	20.80	21.28	4.20
23900	23810	2318	0.752	20.97	21.45	3.66
23931	23837	2322	0.742	20.67	21.15	3.98
23962	23864	2326	0.735	20.47	20.94	4.64
23993	23890	2330	0.743	20.70	21.18	4.66
24024	23917	2334	0.745	20.76	21.24	3.98
24055	23944	2338	0.751	20.94	21.42	3.99
24087	23970	2342	0.752	20.96	21.44	3.92
24118	23996	2346	0.759	21.17	21.66	3.97
24150	24022	2350	0.750	20.92	21.41	4.11
24181	24048	2354	0.742	20.68	21.16	4.53
24213	24074	2358	0.740	20.63	21.10	4.86
24245	24099	2362	0.741	20.64	21.11	4.43
24277	24125	2366	0.749	20.87	21.35	4.20
24309	24150	2370	0.741	20.63	21.11	4.78
24341	24175	2374	0.747	20.83	21.31	4.44
24373	24200	2378	0.743	20.70	21.18	4.99
24405	24225	2382	0.741	20.66	21.13	5.12
24438	24250	2386	0.741	20.63	21.11	5.29
24470	24274	2390	0.736	20.49	20.96	4.86
24503	24298	2394	0.724	20.16	20.62	5.78
24536	24322	2398	0.736	20.51	20.98	6.17
24569	24346	2402	0.743	20.70	21.18	5.21
24602	24370	2406	0.730	20.34	20.80	5.27
24635	24393	2410	0.733	20.41	20.87	4.75
24668	24417	2414	0.742	20.67	21.15	4.94
24701	24440	2418	0.733	20.42	20.88	5.07
24735	24463	2422	0.735	20.48	20.95	4.55
24769	24486	2426	0.735	20.48	20.95	4.93
24802	24508	2430	0.741	20.66	21.14	4.60
24802	24508	2430	0.742	20.69	21.16	4.52
24836	24531	2434	0.742	20.68	21.16	4.82
24870	24553	2438	0.738	20.57	21.04	3.77
24904	24575	2442	0.734	20.44	20.90	5.42
24921	24586	2444	0.733	20.40	20.86	4.64
24956	24607	2448	0.735	20.47	20.94	5.18
24990	24629	2452	0.728	20.26	20.72	5.40
25025	24650	2456	0.730	20.32	20.78	4.71
25060	24671	2460	0.736	20.51	20.98	5.06
25095	24692	2464	0.739	20.60	21.07	4.72
25130	24713	2468	0.735	20.46	20.93	5.91
25165	24733	2472	0.726	20.21	20.68	5.25

25201	24753	2476	0.716	19.93	20.38	6.10
25272	24793	2484	0.714	19.85	20.30	5.03
25308	24812	2488	0.713	19.82	20.27	5.37
25308	24812	2488	0.713	19.83	20.28	5.12
25344	24831	2492	0.726	20.21	20.67	5.25
25381	24850	2496	0.723	20.12	20.58	5.21
25417	24869	2500	0.729	20.29	20.75	6.13
25454	24888	2504	0.732	20.38	20.85	6.17
25491	24906	2508	0.722	20.10	20.56	7.70
25528	24924	2512	0.716	19.92	20.38	5.10
25565	24942	2516	0.715	19.89	20.34	4.49
25602	24959	2520	0.712	19.78	20.23	4.91
25640	24976	2524	0.715	19.90	20.35	5.32
25678	24993	2528	0.720	20.02	20.47	4.32
25716	25010	2532	0.711	19.76	20.20	4.29
25754	25027	2536	0.715	19.88	20.33	5.17
25793	25043	2540	0.722	20.10	20.55	5.33
25831	25059	2544	0.722	20.10	20.56	5.17

GGC 43/CDH 41:

Core ID	Depth (cm)	Age (yrs, Marine13, Clam)	U ^{k'} ₃₇	SST (°C, Prah)	SST (°C, Muller)	C ₃₇ total (nmol/g)
GGC 43	0	841	0.837	23.47	24.03	14.19
GGC 43	2	1032	0.836	23.44	24.00	13.93
GGC 43	2	1032				
GGC 43	4	1220	0.836	23.44	24.00	13.94
GGC 43	4	1220				
GGC 43	6	1405	0.836	23.44	24.00	13.02
GGC 43	8	1587	0.836	23.44	24.00	12.91
GGC 43	10	1766	0.837	23.48	24.04	13.89
GGC 43	10	1766				
GGC 43	12	1943	0.835	23.41	23.97	11.91
GGC 43	14	2117	0.837	23.47	24.03	12.85
GGC 43	16	2289	0.833	23.35	23.91	12.74
GGC 43	18	2458	0.840	23.56	24.12	11.42
GGC 43	20	2624	0.837	23.47	24.03	13.75
GGC 43	22	2788	0.835	23.42	23.97	11.94
GGC 43	24	2950	0.835	23.42	23.98	11.78
GGC 43	26	3109	0.833	23.36	23.91	11.78
GGC 43	28	3266	0.839	23.53	24.09	12.08
GGC 43	28	3266				
GGC 43	30	3420	0.836	23.43	23.99	12.00
GGC 43	32	3572	0.835	23.41	23.97	12.64

GGC 43	34	3722	0.837	23.47	24.03	10.78
GGC 43	36	3870	0.835	23.41	23.97	10.94
GGC 43	38	4015	0.835	23.42	23.98	12.84
GGC 43	40	4158	0.836	23.44	24.00	11.99
GGC 43	42	4299	0.837	23.46	24.02	5.70
GGC 43	42	4299				
GGC 43	42	4299	0.838	23.49	24.05	10.93
GGC 43	44	4438	0.838	23.51	24.07	11.47
GGC 43	44	4438				
GGC 43	46	4575	0.836	23.45	24.01	10.86
GGC 43	48	4710	0.836	23.43	23.99	11.12
GGC 43	50	4843	0.835	23.40	23.96	10.64
GGC 43	52	4974	0.836	23.45	24.01	11.76
GGC 43	52	4974	0.838	23.49	24.06	10.80
GGC 43	54	5104	0.834	23.39	23.94	11.29
GGC 43	54	5104	0.838	23.50	24.06	10.46
GGC 43	56	5231	0.835	23.42	23.98	10.56
GGC 43	58	5356	0.832	23.33	23.89	10.97
GGC 43	60	5480	0.836	23.45	24.01	11.14
GGC 43	62	5601	0.835	23.40	23.96	10.26
GGC 43	64	5721	0.836	23.45	24.01	10.97
GGC 43	64	5721	0.838	23.49	24.05	10.19
GGC 43	66	5839	0.834	23.39	23.95	11.14
GGC 43	68	5956	0.835	23.41	23.96	10.76
GGC 43	70	6070	0.835	23.40	23.96	11.04
GGC 43	72	6184	0.836	23.43	23.98	11.22
GGC 43	74	6295	0.836	23.43	23.99	9.80
GGC 43	76	6405	0.834	23.37	23.92	9.82
GGC 43	78	6513	0.835	23.42	23.98	7.93
GGC 43	80	6620	0.836	23.43	23.99	10.51
GGC 43	82	6725	0.835	23.40	23.96	10.79
GGC 43	82	6725				
GGC 43	84	6828	0.832	23.32	23.88	10.12
GGC 43	86	6930	0.835	23.42	23.97	8.66
GGC 43	88	7031	0.835	23.41	23.97	9.55
GGC 43	90	7130	0.835	23.42	23.98	11.19
GGC 43	92	7228	0.833	23.36	23.91	9.67
GGC 43	94	7324	0.834	23.38	23.93	10.20
GGC 43	96	7419	0.834	23.39	23.94	10.02
GGC 43	98	7513	0.834	23.37	23.93	9.87

GGC 43	100	7605	0.833	23.36	23.92	10.72
GGC 43	102	7696	0.833	23.35	23.91	10.42
GGC 43	104	7786	0.832	23.33	23.88	10.37
GGC 43	106	7874	0.834	23.38	23.93	9.56
GGC 43	106	7874				
GGC 43	108	7962	0.831	23.30	23.86	9.05
GGC 43	110	8048	0.828	23.20	23.75	9.69
GGC 43	112	8133	0.815	22.83	23.37	8.03
GGC 43	112	8133	0.816	22.84	23.39	7.78
GGC 43	114	8216	0.832	23.31	23.87	9.96
GGC 43	116	8299	0.829	23.24	23.79	8.92
GGC 43	118	8380	0.831	23.29	23.85	9.42
GGC 43	118	8380				
GGC 43	120	8461	0.828	23.21	23.76	6.96
GGC 43	122	8540	0.832	23.33	23.89	9.09
GGC 43	124	8618	0.828	23.20	23.75	8.15
GGC 43	126	8695	0.831	23.28	23.84	9.72
GGC 43	128	8771	0.829	23.22	23.78	8.81
GGC 43						
GGC 43	130	8846	0.827	23.18	23.73	9.62
GGC 43	132	8921	0.828	23.20	23.76	9.27
GGC 43	134	8994	0.826	23.15	23.70	8.98
GGC 43	136	9066	0.828	23.21	23.76	8.75
GGC 43						
GGC 43	138	9137	0.831	23.30	23.86	9.53
GGC 43	138	9137				
GGC 43	140	9208	0.829	23.24	23.79	9.60
GGC 43	142	9278	0.829	23.23	23.78	9.51
GGC 43	144	9346	0.829	23.23	23.78	9.55
GGC 43	144	9346				
GGC 43	146	9414	0.828	23.21	23.76	8.37
GGC 43	148	9481	0.823	23.07	23.62	7.19
GGC 43	150	9548	0.829	23.23	23.79	8.39
GGC 43	150	9548				
GGC 43	152	9613	0.825	23.10	23.65	7.56
GGC 43	154	9678	0.827	23.18	23.73	8.39
GGC 43	156	9743	0.830	23.27	23.82	8.68
GGC 43	158	9806	0.827	23.19	23.74	7.92
GGC 43	158	9806				
GGC 43	160	9869	0.826	23.14	23.69	7.50

GGC 43	162	9931	0.832	23.31	23.87	4.78
GGC 43	164	9993	0.826	23.13	23.68	8.15
GGC 43	166	10054	0.825	23.13	23.68	8.02
GGC 43	168	10114	0.829	23.23	23.79	6.76
GGC 43	170	10174	0.823	23.06	23.60	8.07
GGC 43	172	10234	0.823	23.05	23.59	7.60
GGC 43	174	10292	0.822	23.03	23.58	8.00
GGC 43	176	10351	0.808	22.63	23.16	7.31
GGC 43	176	10351				
GGC 43	178	10409	0.811	22.69	23.23	7.76
GGC 43	180	10466	0.824	23.09	23.64	7.32
GGC 43	180	10466				
GGC 43	182	10524	0.820	22.97	23.51	6.68
GGC 43						
GGC 43	184	10580	0.823	23.07	23.62	7.68
GGC 43	186	10637	0.819	22.95	23.49	6.29
GGC 43	188	10693	0.820	22.97	23.52	7.66
GGC 43	190	10749	0.820	22.97	23.51	6.85
GGC 43	192	10804	0.826	23.15	23.70	7.57
GGC 43	194	10860	0.833	23.35	23.91	9.38
GGC 43	194	10860	0.825	23.13	23.68	5.06
GGC 43	196	10915	0.822	23.02	23.57	6.80
GGC 43	196	10915				
GGC 43	198	10970	0.824	23.10	23.65	7.34
GGC 43	200	11025	0.821	22.99	23.54	7.37
GGC 43	202	11079	0.821	23.01	23.56	7.44
GGC 43	204	11134	0.818	22.90	23.44	6.49
GGC 43	206	11189	0.819	22.95	23.49	5.73
GGC 43	208	11243	0.820	22.98	23.52	6.00
GGC 43	210	11298	0.816	22.86	23.40	6.31
GGC 43	212	11352	0.819	22.93	23.48	6.59
GGC 43	214	11407	0.821	22.99	23.53	9.78
GGC 43	216	11462	0.818	22.92	23.46	7.90
GGC 43	218	11517	0.816	22.86	23.40	8.43
GGC 43	220	11572	0.785	21.94	22.46	8.87
GGC 43	222	11627	0.820	22.96	23.50	7.67
GGC 43	224	11682	0.809	22.65	23.18	7.52
GGC 43	226	11738	0.811	22.72	23.26	7.49
GGC 43	226	11738	0.811	22.70	23.24	7.44
GGC 43	228	11794	0.814	22.79	23.33	6.17

GGC 43	230	11851	0.812	22.73	23.27	5.53
GGC 43	232	11908	0.815	22.81	23.35	7.02
GGC 43	234	11965	0.805	22.53	23.06	7.09
GGC 43	236	12022	0.812	22.73	23.27	6.93
GGC 43	236	12022				
GGC 43	238	12081	0.811	22.72	23.25	7.06
GGC 43	240	12139	0.815	22.82	23.36	7.49
GGC 43	242	12199	0.812	22.73	23.27	7.16
GGC 43	244	12259	0.809	22.64	23.18	7.12
GGC 43	246	12319	0.799	22.35	22.87	7.22
GGC 43	248	12381	0.800	22.38	22.91	7.86
GGC 43	248	12381				
GGC 43	250	12443	0.805	22.52	23.06	7.01
GGC 43	252	12506	0.801	22.42	22.95	7.57
GGC 43	254	12570	0.798	22.31	22.84	6.70
GGC 43	256	12634	0.805	22.53	23.06	8.03
GGC 43	258	12700	0.799	22.36	22.89	7.13
GGC 43	260	12766	0.802	22.44	22.97	6.34
GGC 43	262	12834	0.810	22.67	23.20	7.45
GGC 43	264	12903	0.796	22.26	22.78	8.12
GGC 43	266	12973	0.800	22.39	22.92	7.52
GGC 43	268	13044	0.807	22.58	23.11	8.44
GGC 43	270	13116	0.798	22.32	22.84	7.53
GGC 43	272	13189	0.797	22.31	22.83	8.40
GGC 43	274	13264	0.795	22.24	22.76	7.47
GGC 43	276	13340	0.799	22.35	22.88	8.51
GGC 43	278	13418	0.796	22.27	22.80	8.67
GGC 43	280	13497	0.807	22.59	23.12	6.89
GGC 43	282	13578	0.800	22.38	22.90	7.72
GGC 43	284	13661	0.795	22.23	22.76	7.45
GGC 43	286	13745	0.804	22.51	23.04	8.20
GGC 43	286	13745				
GGC 43	288	13831	0.798	22.34	22.86	8.29
GGC 43	290	13918	0.805	22.52	23.05	7.93
GGC 43	292	14008	0.804	22.51	23.04	6.61
GGC 43	294	14099	0.800	22.39	22.91	6.67
GGC 43	296	14193	0.798	22.31	22.84	7.38
GGC 43	298	14288	0.798	22.32	22.85	7.68
GGC 43	300	14386	0.802	22.43	22.96	6.39
GGC 43	302	14486	0.804	22.50	23.03	8.38

GGC 43	304	14588	0.794	22.21	22.73	6.61
GGC 43	306	14692	0.798	22.33	22.85	8.76
GGC 43	308	14799	0.797	22.30	22.83	7.93
GGC 43	310	14908	0.798	22.31	22.84	6.51
GGC 43						
GGC 43	312	15020	0.798	22.33	22.85	7.85
GGC 43	314	15135	0.797	22.30	22.82	8.12
GGC 43	314	15135				
GGC 43	316	15252	0.797	22.29	22.81	8.28
GGC 43						
GGC 43	320	15495	0.802	22.43	22.96	7.03
GGC 43	322	15620	0.800	22.39	22.91	7.91
GGC 43	324	15749	0.797	22.29	22.82	7.17
GGC 43	326	15881	0.797	22.29	22.81	7.49
GGC 43	328	16016	0.799	22.34	22.87	6.11
GGC 43	330	16154	0.795	22.22	22.75	8.08
GGC 43	332	16296	0.789	22.07	22.59	8.61
GGC 43	334	16441	0.800	22.37	22.90	7.15
GGC 43	336	16589	0.792	22.15	22.67	7.78
GGC 43	338	16741	0.795	22.24	22.76	7.85
GGC 43	340	16897	0.789	22.07	22.59	8.28
GGC 43	342	17057	0.779	21.77	22.28	7.44
GGC 43	344	17220	0.787	22.00	22.52	6.57
GGC 43	346	17388	0.777	21.71	22.21	6.52
GGC 43	348	17559	0.780	21.80	22.31	7.78
GGC 43	350	17735	0.776	21.68	22.19	6.40
GGC 43	352	17915	0.786	21.99	22.50	7.77
GGC 43	354	18099	0.769	21.46	21.96	7.92
GGC 43	356	18288	0.785	21.93	22.44	8.39
GGC 43	358	18481	0.779	21.78	22.28	7.12
GGC 43	360	18679	0.795	22.23	22.75	7.95
GGC 43	362	18882	0.790	22.10	22.62	8.00
CDH 41	66	18935	0.795	22.24	22.76	7.56
CDH 41	68	19099	0.799	22.36	22.88	11.12
CDH 41	70	19260	0.794	22.21	22.73	7.73
CDH 41	70	19260				
CDH 41	72	19416	0.795	22.23	22.75	6.96
CDH 41	74	19569	0.797	22.29	22.82	6.65
CDH 41	76	19718	0.793	22.16	22.68	7.46
CDH 41	78	19863	0.795	22.23	22.76	6.31

CDH 41	80	20005	0.798	22.32	22.85	8.03
CDH 41	82	20143	0.792	22.14	22.66	7.74
CDH 41	84	20278	0.794	22.20	22.72	7.58
CDH 41	86	20410	0.788	22.03	22.55	6.69
CDH 41	88	20539	0.790	22.09	22.61	7.04
CDH 41	90	20665	0.791	22.10	22.62	5.83
CDH 41	92	20788	0.797	22.30	22.83	6.39
CDH 41	94	20908	0.797	22.28	22.81	7.74
CDH 41	96	21026	0.799	22.34	22.87	7.06
CDH 41	98	21142	0.800	22.37	22.90	7.19
CDH 41	100	21255	0.795	22.22	22.74	6.69
CDH 41	100	21255				
CDH 41	102	21365	0.794	22.20	22.72	7.21
CDH 41	104	21474	0.794	22.21	22.73	7.17
CDH 41	106	21580	0.796	22.25	22.78	6.13
CDH 41	108	21685	0.793	22.16	22.68	6.61
CDH 41	110	21787	0.788	22.04	22.55	6.94
CDH 41	112	21888	0.790	22.10	22.62	6.05
CDH 41	114	21987	0.780	21.79	22.29	5.84
CDH 41	116	22085	0.791	22.12	22.64	6.02
CDH 41	118	22181	0.798	22.33	22.85	5.72
CDH 41	118	22181				
CDH 41	120	22275	0.791	22.12	22.63	7.18
CDH 41	122	22368	0.798	22.32	22.84	7.45
CDH 41	124	22460	0.797	22.28	22.81	7.62
CDH 41	126	22550	0.799	22.35	22.88	6.96
CDH 41	128	22639	0.794	22.22	22.74	7.44
CDH 41	128	22639				
CDH 41	130	22728	0.793	22.18	22.71	8.43
CDH 41	132	22815	0.796	22.26	22.79	8.16
CDH 41	134	22901	0.798	22.34	22.86	8.37
CDH 41	136	22986	0.798	22.34	22.86	8.14
CDH 41	138	23071	0.796	22.27	22.80	9.21
CDH 41	140	23155	0.790	22.09	22.61	10.55
CDH 41	140	23155				
CDH 41	142	23238	0.791	22.11	22.63	7.32
CDH 41	144	23320	0.789	22.06	22.57	7.79
CDH 41	146	23402	0.794	22.21	22.73	9.25
CDH 41	148	23483	0.801	22.42	22.95	5.45
CDH 41	150	23564	0.800	22.38	22.91	7.45

CDH 41	152	23645	0.800	22.37	22.90	8.27
CDH 41	152	23645				
CDH 41	154	23725	0.799	22.36	22.89	7.63
CDH 41	156	23805	0.803	22.47	23.00	7.19
CDH 41	158	23884	0.804	22.49	23.02	7.73
CDH 41	160	23963	0.802	22.45	22.98	8.08
CDH 41	162	24042	0.801	22.40	22.93	7.72
CDH 41	164	24121	0.800	22.38	22.91	9.02
CDH 41	166	24200	0.804	22.51	23.04	8.39
CDH 41	168	24279	0.801	22.40	22.93	7.27
CDH 41	170	24357	0.804	22.49	23.02	7.99
CDH 41	172	24436	0.799	22.35	22.87	8.44
CDH 41	174	24514	0.797	22.30	22.82	9.97
CDH 41	176	24593	0.796	22.26	22.78	9.90
CDH 41	178	24671	0.796	22.27	22.79	11.41
CDH 41	178	24671				
CDH 41	180	24750	0.801	22.41	22.94	10.84
CDH 41	182	24829	0.801	22.40	22.93	8.91
CDH 41	184	24907	0.808	22.63	23.16	6.16
CDH 41	186	24986	0.796	22.26	22.78	10.08
CDH 41	188	25066	0.798	22.32	22.84	6.81
CDH 41	190	25145	0.803	22.46	22.99	8.56
CDH 41	192	25224	0.805	22.52	23.06	7.46
CDH 41	194	25304	0.803	22.47	23.00	9.88
CDH 41	196	25384	0.804	22.49	23.03	10.17
CDH 41	198	25463	0.807	22.60	23.13	8.29
CDH 41	200	25544	0.805	22.52	23.05	10.76
CDH 41	220	26353	0.800	22.38	22.91	6.90
CDH 41	240	27168	0.805	22.52	23.05	7.84
CDH 41	260	27956	0.809	22.65	23.19	11.98
CDH 41	280	28667	0.810	22.69	23.22	10.05
CDH 41	300	29228	0.802	22.43	22.96	1.60
CDH 41	320	29548	0.802	22.43	22.96	1.40

ME0005A-15MC/17JC:

Core	Depth (cm)	Age (yrs, Marine13, Clam)	U ^{k'} ₃₇	SST (°C, PrahI)	SST (°C, Muller)	C37total (nmol/g)	Brassicasterol (nmol/g)
ME0005-15MC	2.25	6624	0.951	26.84	27.50	0.42	0.11
ME0005-15MC	4.25	7607	0.948	26.74	27.39	0.66	0.15
ME0005-15MC	5.25	8099	0.950	26.80	27.46	0.84	0.17
ME0005-15MC	6.25	8590	0.951	26.83	27.49	0.67	0.13
ME0005-15MC	8.25	9573	0.949	26.76	27.42	0.67	0.18
ME0005-15MC	10.25	10556	0.947	26.71	27.37	0.77	0.15
ME0005-15MC	14.25	12522	0.942	26.57	27.22	0.85	0.14
ME0005-15MC	15.25	13013	0.940	26.49	27.14	0.90	0.14
ME0005-17JC	2.25	7360	0.946	26.67	27.33	0.67	0.12
ME0005-17JC	4.25	8480	0.950	26.79	27.45	0.62	0.09
ME0005-17JC	6.25	9589	0.945	26.64	27.30	0.75	0.12
ME0005-17JC	8.25	10687	0.947	26.70	27.36	0.76	0.18
ME0005-17JC	10.25	11774	0.942	26.57	27.23	0.84	0.12
ME0005-17JC	12.25	12849	0.935	26.35	26.99	1.02	0.16
ME0005-17JC	14.25	13913	0.935	26.36	27.01	1.02	0.14
ME0005-17JC	16.25	14966	0.936	26.37	27.02	0.93	0.14
ME0005-17JC	18.25	16007	0.934	26.33	26.98	1.46	0.18
ME0005-17JC	20.25	17038	0.928	26.16	26.80	1.73	0.26
ME0005-17JC	22.25	18057	0.927	26.11	26.75	2.33	0.28
ME0005-17JC	24.25	19064	0.921	25.96	26.59	2.91	0.31
ME0005-17JC	26.25	20061	0.913	25.72	26.35	3.74	0.52
ME0005-17JC	28.25	21046	0.916	25.80	26.43	3.38	0.44
ME0005-17JC	30.25	22020	0.927	26.12	26.76	4.13	0.46
ME0005-17JC	32.25	22982	0.920	25.91	26.55	4.17	0.42
ME0005-17JC	34.25	23934	0.932	26.27	26.91	4.19	0.35
ME0005-17JC	36.25	24874	0.924	26.02	26.65	3.71	0.38
ME0005-17JC	38.25	25803	0.922	25.97	26.61	3.43	0.39
ME0005-17JC	40.25	26720	0.924	26.02	26.66	3.31	0.32
ME0005-17JC	42.25	27627	0.921	25.94	26.57	2.85	0.28
ME0005-17JC	44.25	28522	0.920	25.92	26.55	2.58	0.33
ME0005-17JC	46.25	29406	0.934	26.33	26.98	2.32	0.25
ME0005-17JC	48.25	30278	0.932	26.26	26.91	1.99	0.23
ME0005-17JC	50.25	31139	0.938	26.45	27.10	1.57	0.18
ME0005-17JC	52.25	31989	0.934	26.33	26.97	1.78	0.17
ME0005-17JC	54.25	32828	0.948	26.75	27.41	1.88	0.11
ME0005-17JC	56.25	33656	0.935	26.36	27.01	1.85	0.15
ME0005-17JC	58.25	34472	0.948	26.73	27.39	1.86	0.11
ME0005-17JC	60.25	35277	0.944	26.61	27.14	1.31	0.10
ME0005-17JC	62.25	36070	0.943	26.58	27.23	1.82	0.16
ME0005-17JC	64.25	36853	0.940	26.49	27.15	1.72	0.13
ME0005-17JC	66.25	37624	0.949	26.77	27.43	2.02	0.12
ME0005-17JC	68.25	38384	0.943	26.60	27.25	2.37	0.13

ME0005-17JC 70.25 39133 0.946 26.68 27.34 2.00 0.11

VNTR01-13GC:

Depth (cm)	Adjusted Depth (cm)	Age (yrs, Marine13, Clam)	U ^k ₃₇	SST (°C, Prah)	SST (°C, Muller)	C37total (nmol/g)	Brassicasterol (nmol/g)
2.25	2.25	764	0.836	23.46	24.01	0.30	0.08
5.25	5.25	2883					0.04
5.75	5.75	3204	0.839	23.53	24.09	0.27	0.09
8.25	8.25	4682					0.10
8.75	8.75	4954	0.841	23.58	24.14	0.25	0.07
10.25	10.25	5727					0.06
10.75	10.75	5971	0.832	23.33	23.89	0.21	0.06
12.25	12.25	6664	0.839	23.52	24.00	0.35	0.07
15.25	15.25	7895					0.07
15.75	15.75	8082	0.838	23.50	24.06	0.22	0.08
17.25	17.25	8615	0.838	23.49	24.06	0.34	0.07
17.75	17.75	8784	0.841	23.60	23.97	0.22	0.08
19.75	19.75	9420	0.829	23.22	23.77		0.08
20.25	20.25	9570	0.855	24.01	24.59	1.26	0.08
23.25	23.25	10401	0.859	24.11	24.69	0.34	0.09
28.25	28.25	11589	0.851	23.88	24.45	0.95	0.10
30.25	30.25	12014	0.845	23.72	24.29	1.37	0.11
32.25	32.25	12419	0.846	23.74	24.31	0.90	0.07
36.25	36.25	13192	0.867	24.34	24.93	1.12	0.06
40.25	40.25	13940	0.824	23.10	23.65	2.09	0.16
44.25	44.25	14690	0.804	22.49	23.02	2.93	0.29
48.25	48.25	15458	0.794	22.20	22.72	3.56	0.36
	50.25	15852			22.83	3.86	
52.25	52.25	16255	0.797	22.31	22.70	3.29	0.33
90.25	54.25	16666	0.793	22.18	22.67	4.48	0.41
92.25	56.25	17086	0.799	22.35	22.69	4.69	0.21
96.25	60.25	17951	0.869	24.40	22.65	5.27	0.24
100.25	64.25	18843	0.797	22.28	22.80	3.96	0.33
102.25	66.25	19298	0.824	23.08	23.63	2.15	0.13
104.25	68.25	19755	0.813	22.77	23.30	2.66	0.23
106.25	70.25	20215	0.816	22.85	23.39	4.10	0.33
108.25	72.25	20675	0.812	22.73	23.27	3.82	0.38
110.25	74.25	21133	0.802	22.43	22.96	4.41	0.33
112.25	76.25	21587	0.815	22.82	23.36	4.01	0.26

114.25	78.25	22036	0.772	21.56	22.06	2.62	0.23
116.25	80.25	22478	0.819	22.95	23.49	5.39	0.19
118.25	82.25	22910	0.810	22.68	23.22	3.15	0.27
120.25	84.25	23330	0.806	22.56	23.10	3.90	0.36
122.25	86.25	23738	0.764	21.32	21.82	3.38	0.54
124.25	88.25	24130	0.754	21.03	21.51	3.93	0.76
126.25	90.25	24505	0.756	21.10	21.58	3.21	0.64
128.25	92.25	24861	0.785	21.95	22.47	4.40	0.47
130.25	94.25	25197	0.799	22.37	22.89	4.80	0.27
132.25	96.25	25512	0.765	21.35	21.85	3.50	0.52
134.25	98.25	25804	0.733	20.41	20.88	3.38	0.70
136.25	100.25	26072	0.758	21.15	21.64	3.57	0.64
138.25	102.25	26316	0.752	20.97	21.46	3.67	0.68
140.25	104.25	26535	0.755	21.07	21.56	4.04	0.61
142.25	106.25	26729	0.745	20.76	21.24	3.72	0.60
144.25	108.25	26898	0.757	21.11	21.60	4.08	0.56
148.25	112.25	27165	0.777	21.71	22.22	1.43	0.44
150.25	114.25	27264	0.799	22.36	22.89	6.56	0.36
154.25	118.25	27403	0.785	21.95	22.47	1.71	0.38
158.25	122.25	27479	0.788	22.03	22.55	0.98	0.21
162.25	126.25	27518	0.795	22.25	22.77	1.85	0.22
166.25	130.25	27556	0.813	22.77	23.30	1.33	0.16
168.25	132.25	27589	0.810	22.69	23.23	0.88	0.17
170.25	134.25	27638	0.836	23.43	23.99	2.02	0.00
190.25	154.25	31016	0.800	22.40	22.92	3.38	0.02

APPENDIX B: Stable Isotope Data

CDH-23:

Depth (cm)	Age (yrs BP, Fairbanks)	$\delta^{13}\text{C}$ (‰, VPDB) <i>U. Peregrina</i>	$\delta^{18}\text{O}$ (‰, VPDB) <i>U. Peregrina</i>
15.5	1105	0.07	1.89
26	1247	0.06	1.90
30	1300	0.17	1.82
42	1457	0.07	1.88
58	1661	0.22	1.78
68.5	1793	0.16	1.75
74	1861	0.14	1.86
89	2043	0.23	1.80
90	2055	0.15	1.85
92.5	2085	0.18	1.81
106	2244	0.15	1.80
118.5	2389	0.09	1.78
138	2609	0.00	1.85
154	2785	0.17	1.80
160	2850	0.12	1.81
180	3063	-0.06	1.88
184.5	3110	0.17	1.87
202.5	3296	0.15	1.79
221.75	3490	0.21	1.83
240	3669	0.12	1.84
249.5	3761	0.19	1.83
260.5	3867	0.24	1.81
280	4050	0.08	1.86
296.5	4203	0.13	1.76
301	4244	0.09	1.84
320.5	4420	0.06	1.76
325.5	4465	0.20	1.80
350.5	4686	0.19	1.76
350.50	4686	0.16	1.80
353.5	4712	0.11	1.75
366	4821	0.10	1.67
382	4958	0.15	1.76
393.5	5056	0.09	1.73
400	5112	0.20	1.83
415.5	5242	0.17	1.85

421.5	5292	0.16	1.76
442.5	5467	-0.01	1.82
444.5	5483	0.02	1.81
460.5	5615	0.17	1.76
500	5937	0.29	1.91
520	6098	0.05	1.78
550	6339	0.28	1.89
571.5	6512	-0.01	1.83
600	6739	0.11	1.85
603.5	6767	0.12	1.79
660.5	7222	0.02	1.77
675	7338	0.05	1.84
702.5	7559	0.08	1.95
718	7683	0.23	1.99
739	7853	-0.10	1.82
760.5	8027	-0.27	1.99
781	8193	0.03	1.96
791.5	8278	-0.34	1.95
809	8421	-0.12	1.98
823.5	8540	-0.19	2.01
825.5	8556	-0.11	1.95
850.25	8760	-0.36	1.89
873.5	8951	-0.10	1.86
893.5	9117	-0.22	1.84
900	9170	-0.36	1.96
916.5	9307	-0.42	1.87
925	9378	-0.24	1.91
940.5	9507	-0.38	1.92
967.5	9732	-0.28	1.89
975	9794	-0.30	2.07
1000.25	10005	-0.34	1.91
1006.5	10057	-0.28	1.92
1025	10211	-0.33	2.09
1050	10419	-0.39	2.11
1062.5	10523	-0.29	2.01
1073.5	10614	-0.41	2.03
1075	10627	-0.41	1.96
1090	10751	-0.28	2.04
1100	10833	-0.29	2.36
1120	10998	-0.39	2.23

1125	11038	-0.49	2.27
1135	11120	-0.47	2.21
1150	11242	-0.55	2.52
1160	11323	-0.59	2.29
1175	11443	-0.68	2.41
1190	11563	-0.56	2.46
1200	11642	-0.41	2.24
1220.5	11802	-0.55	2.36
1230	11876	-0.59	2.46
1250	12029	-0.46	2.50
1275	12217	-0.37	2.32
1285	12291	-0.68	2.62
1297	12378	-0.79	2.51
1302	12414	-0.62	2.69
1325	12578	-0.49	2.68
1340	12682	-0.67	2.71
1350	12750	-0.43	2.80
1355	12783	-0.46	2.48
1358.5	12806	-0.73	2.77
1360.5	12820	-0.51	2.35
1375	12915	-0.65	2.48
1400	13072	-0.68	2.86
1435.5	13282	-0.56	2.76
1440	13308	-0.49	2.42
1470.5	13471	-0.48	2.48
1475	13494	-0.59	2.50
1486.5	13551	-0.77	2.44
1489.5	13565	-0.80	2.63
1500	13615	-0.60	2.69
1525	13724	-0.58	2.59
1550	13821	-0.89	2.76
1560	13856	-1.03	2.78
1575	13905	-0.86	2.93
1585	13935	-0.88	2.68
1600	13975	-0.68	2.56
1625	14029	-0.97	3.02
1645.5	14062	-0.80	2.38
1651	14069	-0.85	2.49
1675	14089	-0.56	2.43
1700	14092	-0.90	2.40

1705.5	14091	-0.71	2.41
1712.5	14087	-0.70	2.38
1713	14087	-0.63	2.54

GGC 43/CDH 41:

Core	Depth (cm)	Age (yrs BP, Fairbanks)	$\delta^{13}\text{C}$ (‰, VPDB) <i>U. Peregrina</i>	$\delta^{18}\text{O}$ (‰, VPDB) <i>U. Peregrina</i>
GGC43	2	1340	-0.59	2.134
GGC43	20	3178	-0.67	2.461
GGC43	40	4725	-0.70	2.159
GGC43	60	5942	-0.58	2.136
GGC43	80	6969	-0.60	2.104
GGC43	100	7896	-0.63	2.099
GGC43	120	8771	-0.77	2.250
GGC43	131	9216	-0.78	2.205
GGC43	140	9611	-0.70	2.224
GGC43	146	9836	-0.66	2.272
GGC43	160	10413	-0.71	2.353
GGC43	171	10814	-0.79	2.130
GGC43	176	10998	-0.77	2.190
GGC43	180	11161	-0.69	2.398
GGC43	195	11675	-0.82	2.448
GGC43	200	11836	-0.83	2.367
GGC43	203	11931	-0.65	2.421
GGC43	205	11993	-0.87	2.494
GGC43	210	12143	-0.75	2.581
GGC43	215	12289	-0.81	2.805
GGC43	220	12429	-0.75	2.879
GGC43	225	12565	-0.83	2.778
GGC43	230	12696	-0.82	2.708
GGC43	235	12823	-0.83	2.945
GGC43	240	12946	-0.73	2.866
GGC43	245	13066	-0.80	2.939
GGC43	250	13185	-0.89	2.762
GGC43	255	13302	-0.83	2.632
GGC43	260	13420	-0.76	2.880
GGC43	265	13539	-0.88	2.761
GGC43	270	13661	-0.64	3.137
GGC43	275	13787	-0.81	2.863

GGC43	280	13920	-0.83	2.682
GGC43	285	14062	-0.82	2.930
GGC43	290	14215	-0.80	2.904
GGC43	295	14381	-0.80	2.996
GGC43	300	14564	-0.73	3.034
GGC43	305	14766	-0.80	3.014
GGC43	310	14991	-0.78	3.047
GGC43	313	15138	-0.61	3.144
GGC43	315	15241	-0.77	3.078
GGC43	318	15406	-0.75	3.023
GGC43	320	15522	-0.65	3.249
GGC43	323	15707	-0.60	3.038
GGC43	325	15837	-0.80	2.863
GGC43	328	16044	-0.66	3.030
GGC43	330	16191	-0.71	3.040
GGC43	333	16423	-0.64	2.980
GGC43	335	16587	-0.65	3.195
GGC43	338	16848	-0.57	3.206
GGC43	340	17032	-0.82	3.379
GGC43	343	17324	-0.55	3.172
GGC43	345	17531	-0.83	3.156
GGC43	348	17858	-0.65	3.188
GGC43	350	18088	-0.75	3.334
GGC43	353	18454	-0.44	3.252
GGC43	355	18711	-0.84	3.218
GGC43	358	19119	-0.48	3.205
GGC43	360	19406	-0.71	3.132
GGC43	363	19859	-0.58	3.088
GGC43	365	20178	-0.71	3.318
CDH 41	85	20265	-0.67	3.454
CDH 41	87	20431	-0.45	3.431
CDH 41	90	20654	-0.75	3.164
CDH 41	92	20830	-0.41	3.266
CDH 41	95	21059	-0.74	3.254
CDH 41	97	21207	-0.58	3.195
CDH 41	100	21422	-0.70	3.128
CDH 41	105	21762	-0.67	3.404
CDH 41	110	22064	-0.74	3.337
CDH 41	115	22376	-0.77	3.453
CDH 41	120	22652	-0.68	3.123

CDH 41	125	22897	-0.72	3.327
CDH 41	130	23137	-0.76	3.265
CDH 41	135	23362	-0.60	3.348
CDH 41	140	23583	-0.67	3.316
CDH 41	145	23772	-0.69	3.248
CDH 41	150	23962	-0.66	3.143
CDH 41	155	24144	-0.54	3.237
CDH 41	160	24321	-0.60	3.073
CDH 41	165	24494	-0.69	3.266
CDH 41	170	24666	-0.61	3.041
CDH 41	175	24840	-0.72	3.280
CDH 41	180	25026	-0.58	3.233
CDH 41	185	25198	-0.63	3.290
CDH 41	190	25387	-0.62	3.237
CDH 41	195	25584	-0.69	3.283
CDH 41	200	25801	-0.64	3.200

CDH-26

Depth (cm)	Age (yrs BP)	$\delta^{13}\text{C}$ (‰, VPDB) <i>U. Peregrina</i>	$\delta^{18}\text{O}$ (‰, VPDB) <i>U. Peregrina</i>
32	1924	-0.29	2.96
64	2976	-0.39	3.06
96	3948	-0.29	2.95
128	4846	-0.52	2.89
171.5	5956	-0.37	2.88
203.5	6698	-0.41	2.91
235.5	7382	-0.49	2.92
267.5	8012	-0.57	2.99
325	9022	-0.57	3.10
357	9523	-0.38	3.21
389	9985	-0.63	3.12
410	10268	-0.79	3.27
421	10411	-0.60	3.13
478	11093	-0.64	3.29
510	11437	-0.61	3.36
542	11756	-0.78	3.14
574	12054	-0.77	3.40
631	12539	-0.63	3.58
663	12788	-0.86	3.71

695	13025	-0.99	3.70
727	13250	-1.07	3.76
784.5	13631	-1.04	3.78
816.5	13832	-1.10	3.77
848.5	14028	-1.35	3.82
880.5	14220	-0.82	3.81
959.5	14680	-0.77	3.83
971.5	14749	-0.80	3.81
1003.5	14933	-0.86	3.88
1035.5	15117	-0.88	3.94
1092.5	15446	-0.87	3.94
1124.5	15633	-0.66	3.99
1156.5	15822	-0.92	4.05
1188.5	16013	-1.26	4.13
1247.5	16371	-0.89	4.19
1263.5	16470	-1.24	4.19
1295.5	16670	-1.46	4.25
1343.5	16975	-0.96	4.38
1400.5	17345	-0.89	4.34
1432.5	17557	-0.97	4.40
1464.5	17771	-1.43	4.43
1496.5	17988	-1.21	4.50
1555.5	18394	-1.47	4.45
1587.5	18616	-1.29	4.51
1619.5	18841	-1.46	4.50
1651.5	19066	-0.98	4.40
1709	19475	-1.13	4.46
1741	19703	-1.17	4.48
1773	19933	-0.89	4.51
1805	20162	-1.61	4.56
1863	20579	-0.97	4.39
1895	20809	-1.08	4.56
1927	21039	-0.90	4.56
1959	21269	-1.03	4.42
2016	21679	-0.96	4.44
2048	21909	-0.71	4.44
2080	22140	-1.21	4.44
2120	22429	-0.45	4.40
2169	22785	-0.98	4.50
2201	23020	-0.51	4.36

2233	23256	-0.63	4.36
2265	23495	-0.70	4.42
2322	23931	-0.75	4.32
2354	24181	-0.75	4.31
2386	24438	-0.69	4.40
2418	24701	-0.67	4.34
2475	25192	-0.89	4.38
2507	25481	-0.59	4.49
2539	25783	-0.61	4.42

Individual Benthic Stable Isotope Data

Core ID	Depth	Midpoint Age (yrs)	Weight (mg)	$\delta^{13}\text{C}$ (‰, VPDB) <i>U. Peregrina</i>	$\delta^{18}\text{O}$ (‰, VPDB) <i>U. Peregrina</i>
MC-18A	0.5	49	40	-0.368	2.901
MC-18A	0.5	49	40	-0.458	2.837
MC-18A	1.5	49	30	-0.360	2.847
MC-18A	1.5	49	42	-0.453	2.943
MC-18A	2.5	49	32	-0.299	2.921
MC-18A	2.5	49	28	-0.417	2.757
MC-18A	3.5	49	34	-0.232	2.954
MC-18A	3.5	49	44	-0.214	2.833
MC-18A	4.5	49	34	-0.517	2.692
MC-18A	4.5	49	42	-0.377	2.838
MC-18A	5.5	49	24	-0.106	3.090
MC-18A	5.5	49	36	-0.287	2.898
MC-18A	6.5	49	30	-0.289	2.829
MC-18A	6.5	49	36	-0.450	2.819
MC-18A	7.5	49	28	-0.636	2.954
MC-18A	7.5	49	46	-0.131	2.893
MC-18A	8.5	49	32	-0.304	2.943
MC-18A	8.5	49	38	-0.522	2.827
MC-18A	9.5	49	30	-0.389	2.788
MC-18A	9.5	49	34	-0.233	2.678
MC-18A	10.5	49	24	-0.431	3.108
MC-18A	10.5	49	36	-0.244	2.930
MC-18A	10.5	49	34	-0.423	2.936
MC-18A	11.5	49	32	-0.329	3.070
MC-18A	11.5	49	36	-0.457	2.815
MC-18A	12.5	49	26	-0.630	2.863

MC-18A	12.5	49	42	-0.331	2.866
MC-18A	13.5	49	30	-0.304	2.853
MC-18A	13.5	49	28	-0.256	2.682
MC-18A	14.5	49	32	-0.262	2.676
MC-18A	14.5	49	32	-0.246	2.813
MC-18A	15.5	49	40	-0.374	2.799
MC-18A	15.5	49	34	-7.634	-5.099
MC-18A	16.5	49	34	-0.205	2.831
MC-18A	16.5	49	42	-0.296	2.895
MC-18A	17.5	49	28	-0.475	2.882
MC-18A	17.5	49	32	-0.081	2.829
MC-18A	18.5	49	32	-0.380	2.829
MC-18A	18.5	49	38	-0.295	2.852
MC-18A	19.5	49	40	-0.181	2.822
MC-18A	19.5	49	42	-0.314	2.879
CDH-26	71	3267	34	0.082	2.916
CDH-26	71	3267	30	-0.424	2.820
CDH-26	71	3267	28	-0.159	2.857
CDH-26	71	3267	36	-0.283	2.809
CDH-26	71	3267	34	-0.455	2.847
CDH-26	71	3267	36	-0.417	2.847
CDH-26	72	3267	20	-0.629	2.938
CDH-26	72	3267	36	-0.374	2.834
CDH-26	72	3267	38	-0.565	2.849
CDH-26	72	3267	30	-0.347	2.904
CDH-26	72	3267	36	-0.395	2.784
CDH-26	72	3267	48	-0.273	2.842
CDH-26	73	3267	24	-0.077	2.945
CDH-26	73	3267	34	-0.342	3.101
CDH-26	73	3267	34	-0.173	2.979
CDH-26	73	3267	38	-0.412	2.761
CDH-26	73	3267	28	-0.430	2.788
CDH-26	73	3267	32	-0.575	2.735
CDH-26	73	3267	40	-0.223	2.729
CDH-26	74	3267	24	-0.598	3.081
CDH-26	74	3267	28	-0.787	2.849
CDH-26	74	3267	24	-0.422	2.684
CDH-26	74	3267	42	-0.213	2.874
CDH-26	74	3267	32	-0.197	2.696
CDH-26	74	3267	52	-0.186	2.886

CDH-26	74	3267	62	-0.116	2.742
CDH-26	75	3267	24	-0.075	2.626
CDH-26	75	3267	26	-0.617	2.852
CDH-26	75	3267	38	-0.331	2.832
CDH-26	75	3267	34	-0.307	2.929
CDH-26	75	3267	36	-0.196	2.748
CDH-26	75	3267	46	-0.208	2.814
CDH-26	76	3267	26	-0.040	3.062
CDH-26	76	3267	26	-0.276	2.889
CDH-26	76	3267	24	-0.397	2.978
CDH-26	76	3267	28	-0.351	2.933
CDH-26	76	3267	24	-0.886	2.762
CDH-26	76	3267	56	-0.511	2.886
CDH-26	76	3267	30	-0.469	2.848
CDH-26	76	3267	46	-0.304	2.773
CDH-26	103.5	4080	36	-0.103	2.917
CDH-26	103.5	4080	22	-0.167	2.503
CDH-26	103.5	4080	38	-0.217	2.961
CDH-26	103.5	4080	28	-0.462	2.905
CDH-26	103.5	4080	36	-0.289	2.675
CDH-26	103.5	4080	42	-0.351	2.826
CDH-26	104.5	4080	22	-0.188	3.109
CDH-26	104.5	4080	26	-0.631	2.939
CDH-26	104.5	4080	38	-0.120	2.648
CDH-26	104.5	4080	34	-0.076	2.609
CDH-26	104.5	4080	32	-0.194	2.825
CDH-26	104.5	4080	56	-0.267	2.786
CDH-26	105.5	4080	24	-0.254	2.976
CDH-26	105.5	4080	32	-0.443	2.865
CDH-26	105.5	4080	28	-0.390	2.802
CDH-26	105.5	4080	34	-0.364	2.906
CDH-26	105.5	4080	34	-0.154	2.844
CDH-26	106.5	4080	34	-0.759	2.895
CDH-26	106.5	4080	24	0.027	3.070
CDH-26	106.5	4080	34	-0.201	3.043
CDH-26	106.5	4080	26	-0.091	2.752
CDH-26	106.5	4080	38	-0.174	2.877
CDH-26	106.5	4080	34	-0.396	2.771
CDH-26	106.5	4080	32	-0.472	2.788
CDH-26	106.5	4080	32	-0.471	2.778

CDH-26	107.5	4080	34	-0.253	3.067
CDH-26	107.5	4080	16	-0.665	2.889
CDH-26	107.5	4080	28	-0.102	2.563
CDH-26	107.5	4080	32	-0.265	2.650
CDH-26	107.5	4080	40	-0.482	2.840
CDH-26	107.5	4080	50	-0.480	2.831
CDH-26	108.5	4080	32	-0.169	2.849
CDH-26	108.5	4080	36	-0.124	2.691
CDH-26	108.5	4080	46	-0.307	2.697
CDH-26	108.5	4080	42	-0.430	2.831
CDH-26	108.5	4080	50	-0.334	2.836
CDH-26	109.5	4080	42	-0.484	2.123
CDH-26	109.5	4080	42	-0.765	2.285
CDH-26	109.5	4080	38	-0.512	2.914
CDH-26	109.5	4080	36	-0.482	2.883
CDH-26	109.5	4080	50	-0.304	2.812
CDH-26	198.5	6194	44	-0.495	2.811
CDH-26	198.5	6194	38	-0.353	2.915
CDH-26	198.5	6194	56	-0.419	2.811
CDH-26	198.5	6194	36	-0.537	2.820
CDH-26	199.5	6194	38	-0.275	2.792
CDH-26	199.5	6194	34	-0.563	2.745
CDH-26	199.5	6194	38	-0.429	2.788
CDH-26	199.5	6194	44	-0.384	2.768
CDH-26	200.5	6194	24	-0.392	3.023
CDH-26	200.5	6194	30	-0.567	2.799
CDH-26	200.5	6194	38	-0.433	2.808
CDH-26	200.5	6194	40	-0.571	2.692
CDH-26	201.5	6194	34	-0.570	2.615
CDH-26	201.5	6194	40	-0.491	2.746
CDH-26	201.5	6194	58	-0.532	2.722
CDH-26	201.5	6194	48	-0.220	2.725
CDH-26	202.5	6194	48	-0.078	3.215
CDH-26	202.5	6194	18	-0.461	2.876
CDH-26	202.5	6194	34	-0.708	2.749
CDH-26	202.5	6194	40	-0.644	2.707
CDH-26	202.5	6194	34	-0.358	2.747
CDH-26	203.5	6194	32	-0.474	2.763
CDH-26	203.5	6194	44	-0.397	2.710
CDH-26	203.5	6194	40	-0.192	2.813

CDH-26	203.5	6194	50	-0.470	2.692
CDH-26	204.5	6194	28	-0.524	2.806
CDH-26	204.5	6194	34	-0.406	2.832
CDH-26	204.5	6194	34	-0.597	2.795
CDH-26	204.5	6194	50	-0.288	2.612
CDH-26	205.5	6194	34	-0.443	2.787
CDH-26	205.5	6194	38	-0.670	2.697
CDH-26	205.5	6194	34	-0.476	2.811
CDH-26	205.5	6194	44	-0.390	2.747
CDH-26	206.5	6194	22	-0.492	2.726
CDH-26	206.5	6194	28	-0.412	2.836
CDH-26	206.5	6194	38	-0.461	2.716
CDH-26	206.5	6194	46	-0.553	2.691
CDH-26	207.5	6194	26	-0.608	2.858
CDH-26	207.5	6194	26	-0.563	2.797
CDH-26	207.5	6194	30	-0.624	2.823
CDH-26	207.5	6194	34	-0.308	2.773
CDH-26	296.5	8004	30	-0.370	3.072
CDH-26	296.5	8004	34	-0.575	2.845
CDH-26	296.5	8004	32	-0.544	2.652
CDH-26	297.5	8004	28	-0.855	2.927
CDH-26	297.5	8004	28	-0.531	2.909
CDH-26	297.5	8004	40	-0.554	2.821
CDH-26	298.5	8004	26	-0.624	2.878
CDH-26	298.5	8004	32	-0.742	2.836
CDH-26	298.5	8004	28	-0.543	2.864
CDH-26	299.5	8004	20	-0.562	3.082
CDH-26	299.5	8004	24	-0.860	2.837
CDH-26	299.5	8004	36	-0.311	2.855
CDH-26	300.5	8004	34	-0.720	2.861
CDH-26	300.5	8004	48	-0.666	2.932
CDH-26	300.5	8004	44	-0.892	2.922
CDH-26	300.5	8004	74	-0.772	2.802
CDH-26	301.5	8004	32	-0.527	2.969
CDH-26	301.5	8004	26	-0.589	2.968
CDH-26	301.5	8004	32	-0.827	2.792
CDH-26	301.5	8004	44	-0.513	2.849
CDH-26	302.5	8004	30	-1.010	2.333
CDH-26	302.5	8004	42	-0.429	2.970
CDH-26	302.5	8004	46	-0.658	2.832

CDH-26	302.5	8004	36	-0.541	2.839
CDH-26	303.5	8004	34	-0.724	2.796
CDH-26	303.5	8004	30	-0.584	2.846
CDH-26	303.5	8004	38	-0.483	2.782
CDH-26	303.5	8004	44	-0.511	2.703
CDH-26	304.5	8004	24	-0.545	2.893
CDH-26	304.5	8004	38	-0.694	2.985
CDH-26	304.5	8004	30	-0.669	2.969
CDH-26	304.5	8004	38	-0.283	2.925
CDH-26	305.5	8004	32	-0.757	3.023
CDH-26	305.5	8004	32	-0.642	2.977
CDH-26	305.5	8004	26	-0.923	2.456
CDH-26	306.5	8004	32	-0.628	2.828
CDH-26	306.5	8004	76	-0.527	3.121
CDH-26	306.5	8004	36	-0.443	2.861
CDH-26	307.5	8004	30	-0.835	2.783
CDH-26	307.5	8004	36	-0.499	2.802
CDH-26	307.5	8004	40	-0.566	2.860
CDH-26	79.5	3493	44	-0.462	2.863
CDH-26	79.5	3493	47	-0.183	2.910
CDH-26	79.5	3493	23	0.020	2.602
CDH-26	79.5	3493	26	-0.339	2.846
CDH-26	81.5	3493	50	-0.306	2.841
CDH-26	79.5	3493	44	-0.130	2.793
CDH-26	83.5	3493	25	-0.164	2.888
CDH-26	81.5	3493	45	0.013	2.833
CDH-26	81.5	3493	29	0.043	2.874
CDH-26	79.5	3493	26	-0.140	2.838
CDH-26	81.5	3493	29	-0.155	3.094
CDH-26	81.5	3493	30	-0.256	2.951
CDH-26	84.5	3493	61	-0.317	2.856
CDH-26	84.5	3493	23	-0.427	2.775
CDH-26	84.5	3493	39	-0.269	2.799
CDH-26	84.5	3493	48	-0.203	2.906
CDH-26	85.5	3493	46	-0.216	2.829
CDH-26	84.5	3493	28	0.101	3.355
CDH-26	85.5	3493	43	-0.446	2.858
CDH-26	84.5	3493	25	-0.334	3.087
CDH-26	85.5	3493	36	-0.534	2.777
CDH-26	85.5	3493	59	-0.256	2.811

CDH-26	85.5	3493	32	-0.391	2.905
CDH-26	80.5	3493	35	-0.496	2.699
CDH-26	80.5	3493	28	-0.409	2.894
CDH-26	80.5	3493	40	-0.349	2.666
CDH-26	80.5	3493	37	-0.148	3.054
CDH-26	80.5	3493	58	-0.522	2.824
CDH-26	82.5	3493	38	-0.621	2.873
CDH-26	82.5	3493	25	-0.373	2.861
CDH-26	82.5	3493	17	0.234	3.090
CDH-26	82.5	3493	38	-0.238	2.934
CDH-26	82.5	3493	27	-0.433	3.027
CDH-26	82.5	3493	28	-0.118	2.941
CDH-26	83.5	3493	38	-0.347	2.753
CDH-26	83.5	3493	36	-0.336	2.674
CDH-26	83.5	3493	31	-0.340	2.829
CDH-26	81.5	3493	87	-0.078	3.047
CDH-26	83.5	3493	41	-0.367	2.840
CDH-26	83.5	3493	33	-0.476	2.818
CDH-26	240.5	7003	26	-0.349	2.899
CDH-26	240.5	7003	45	-0.579	2.876
CDH-26	240.5	7003	41	-0.457	2.836
CDH-26	240.5	7003	24	-0.353	2.843
CDH-26	241.5	7003	26	-0.538	2.972
CDH-26	241.5	7003	37	-0.423	2.907
CDH-26	241.5	7003	30	-0.394	2.826
CDH-26	241.5	7003	23	-0.366	2.930
CDH-26	243.5	7003	28	-0.643	2.811
CDH-26	243.5	7003	30	-0.544	2.838
CDH-26	243.5	7003	36	-0.269	2.816
CDH-26	243.5	7003	24	-0.558	2.874
CDH-26	242.5	7003	30	-0.400	2.823
CDH-26	242.5	7003	24	-0.685	2.837
CDH-26	242.5	7003	34	-0.828	2.735
CDH-26	248.5	7003	38	-0.865	3.172
CDH-26	248.5	7003	46	-0.556	2.774
CDH-26	248.5	7003	31	-1.012	1.227
CDH-26	249.5	7003	51	-0.665	3.065
CDH-26	249.5	7003	36	-0.415	2.717
CDH-26	249.5	7003	24	-0.426	2.891
CDH-26	249.5	7003	30	-0.409	3.062

CDH-26	244.5	7003	32	-0.355	2.916
CDH-26	244.5	7003	42	-0.283	2.863
CDH-26	244.5	7003	35	-0.362	2.852
CDH-26	244.5	7003	32	-0.356	2.894
CDH-26	247.5	7003	41	-0.559	2.947
CDH-26	247.5	7003	40	-0.749	2.937
CDH-26	247.5	7003	39	-0.591	2.818
CDH-26	247.5	7003	34	-0.292	2.803
CDH-26	245.5	7003	35	-0.388	2.889
CDH-26	245.5	7003	33	-0.534	2.812
CDH-26	245.5	7003	37	-0.397	3.186
CDH-26	245.5	7003	37	-0.388	3.148
CDH-26	245.5	7003	40	-0.524	2.878
CDH-26	246.5	7003	55	-0.633	2.787
CDH-26	246.5	7003	33	-0.424	3.225
CDH-26	246.5	7003	36	-0.820	2.779
CDH-26	246.5	7003	58	-0.598	2.845

Small Standards

Standard: BYM63150 ($\delta^{13}\text{C} = -2.26\text{‰}$, $\delta^{18}\text{O} = -6.52\text{‰}$)

Weight (mg)	$\delta^{13}\text{C}$ (‰, VPDB) <i>U. Peregrina</i>	$\delta^{18}\text{O}$ (‰, VPDB) <i>U. Peregrina</i>
36	-2.670	-7.131
12	-2.399	-6.734
12	-2.487	-6.488
16	-2.335	-6.519
28	-2.326	-6.365
35	-2.252	-6.184
18	-2.285	-6.389
16	-2.277	-6.365
18	-2.388	-6.502
?	-2.331	-6.613
26	-2.346	-6.350
22	-2.311	-6.400
22	-2.305	-6.452
33	-2.204	-6.314
14	-2.281	-6.350
24	-2.274	-6.491
22	-2.269	-6.398

22	-2.311	-6.420
22	-2.352	-6.348
22	-2.266	-6.317
28	-2.293	-6.363
24	-2.249	-6.232
35	-2.338	-6.390
30	-2.291	-6.312
22	-2.294	-6.292
31	-2.266	-6.243
36	-2.245	-6.271
26	-2.330	-6.393
70	-2.314	-6.485
70	-2.314	-6.485
32	-2.233	-6.251
28	-2.321	-6.396
50	-2.263	-6.445
50	-2.263	-6.445
26	-2.304	-6.436
28	-2.272	-6.347
30	-2.253	-6.268
34	-2.287	-6.396
33	-2.299	-6.331
36	-2.473	-6.553
30	-2.304	-6.241
38	-2.351	-6.490
38	-2.342	-6.449
26	-2.320	-6.445
28	-2.324	-6.438
30	-2.296	-6.493
32	-2.299	-6.504
33	-2.330	-6.456
26	-2.279	-6.352
43	-2.338	-6.486
35	-2.288	-6.406
33	-2.359	-6.504
35	-2.342	-6.482
35	-2.372	-6.455
35	-2.472	-6.684
32	-2.326	-6.397
34	-2.306	-6.481

70	-2.303	-6.605
29	-2.341	-6.514
33	-2.305	-6.446
35	-2.304	-6.380
35	-2.323	-6.540
40	-2.368	-6.576
33	-2.342	-6.501
38	-2.349	-6.367
32	-2.323	-6.465
36	-2.354	-6.588
36	-2.365	-6.534
70	-2.328	-6.508
37	-2.344	-6.474
37	-2.274	-6.473
70	-2.311	-6.473
80	-2.292	-6.503
34	-2.321	-6.428
32	-2.329	-6.483
38	-2.353	-6.502
70	-2.303	-6.605
70	-2.303	-6.605
36	-2.328	-6.513
32	-2.362	-6.654
110	-2.291	-6.429
33	-2.322	-6.489
50	-2.263	-6.445
?	-2.331	-6.613
?	-2.331	-6.613
90	-2.302	-6.451
40	-2.314	-6.537
70	-2.297	-6.510
54	-2.351	-6.512
57	-2.058	-6.465
70	-2.312	-6.368
83	-2.362	-6.626
59	-2.341	-6.512
70	-2.328	-6.508
70	-2.328	-6.508
40	-2.314	-6.537
35	-2.369	-6.570

64	-2.332	-6.635
70	-2.333	-6.499
70	-2.312	-6.368
31	-2.351	-6.602
58	-2.314	-6.473

APPENDIX C: Radiocarbon data

Core	Depth in Core	Weight (mg)	Benthics	Planktonics	Accession Number	14C age (yrs B.P.)	14C Error (years)	Outlier*
CDH 23	30	2.8	<i>Uvigerina</i>			1960	20	
CDH 23	30	4		N. dutertrei	OS-77623	1680	40	
CDH 23	160	2.3	<i>Uvigerina</i>			3330	25	
CDH 23	160	3		N. dutertrei	OS-77579	3420	35	
CDH 23	260.5	2.6	<i>mixed benthics</i>			4290	25	
CDH 23	260.5	4		N.dutertrei	OS-77576	3980	35	
CDH 23	350.5	4.2	<i>Uvigerina</i>			4730	20	
CDH 23	350.5	3.3	<i>H. concentrica</i>			4660	25	
CDH 23	350.5	3		N.dutertrei	OS-77575	4540	40	
CDH 23	400	5		N. dutertrei	OS-84123	4800	35	
CDH 23	460.5	5		N. dutertrei	OS- 77546	5290	50	
CDH 23	500	2.6	<i>mixed benthics</i>			6370	50	
CDH 23	500	5		N. dutertrei	OS-84122	5720	30	
CDH 23	550	3.1	<i>mixed benthics</i>			6380	50	
CDH 23	550	5		N. dutertrei	OS-84108	6230	35	
CDH 23	600	2.2	<i>mixed benthics</i>			6850	25	
CDH 23	600	4		N. dutertrei	OS-84168	6510	45	
CDH 23	660.5	4		N. dutertrei	OS- 77574	6900	40	
CDH 23	781	3		N. dutertrei	OS-84121	7950	40	
CDH 23	850.25	4		N. dutertrei	OS- 77580	8640	45	
CDH 23	900	7.5	<i>mixed benthics</i>			8990	30	
CDH 23	900	4		N. dutertrei	OS-84120	8600	45	
CDH 23	1000.25	6.6	<i>mixed benthics</i>			9510	30	
CDH 23	1000.25	7		N. dutertrei	OS- 77547	9220	60	
CDH 23	1100	5.2	<i>mixed benthics</i>			10150	35	
CDH 23	1100	7		N. dutertrei	OS-84115	9920	40	
CDH 23	1230	7.6	<i>H. concentrica</i>			10950	45	
CDH 23	1230	4		N. dutertrei	OS-84130	10750	45	
CDH 23	1285	4		N. dutertrei	OS-84124	10900	50	
CDH 23	1360.5	5.4	<i>Bolivina</i>			11500	40	
CDH 23	1360.5	6.9	<i>H. concentrica</i>			11650	35	
CDH 23	1360.5	6		N. dutertrei	OS- 77573	11650	55	
CDH 23	1440	4	<i>Bolivina</i>			12200	45	B>P
CDH 23	1440	10		N. dutertrei	OS-84155	12000	55	
CDH 23	1500	5.6	<i>mixed benthics</i>			12350	40	
CDH 23	1500	5		N. dutertrei	OS- 77544	12200	70	
CDH 23	1600	10		N. dutertrei	OS-84116	12600	55	

CDH 23	1700	6.4	<i>Bolivina</i>			12950	45	
CDH 23	1700	4		<i>N. dutertrei</i>	OS- 77543	12800	95	
CDH 23	118-119	5.8		<i>N. dutertrei</i>	OS-101834	3000	20	
CDH 23	42-43	6.5		<i>N. dutertrei</i>	OS-101833	2110	20	
GGC43	2	2.4	<i>Uvigerina</i>		OS-109398	2080	20	
GGC43	2	3.1		<i>G. ruber</i>	OS-99077	1990	50	
GGC43	40	3.5	<i>Uvigerina</i>		OS-109328	5420	30	
GGC43	40	3.9		<i>G. ruber</i>	OS-109329	4630	40	
GGC43	70.5	1.9	<i>Uvigerina</i>		OS-109456	6520	30	
GGC43	70.5	4.5		<i>G. ruber</i>	OS-109330	6050	35	
GGC43	100	3.8	<i>Uvigerina</i>		OS-109331	8180	40	
GGC43	100	3		<i>G. ruber</i>	OS-99078	8030	50	
GGC43	140	3.5	<i>Uvigerina</i>		OS-109332	9040	40	
GGC43	140	4.3		<i>G. ruber</i>	OS-109333	8800	40	
GGC43	180	3	<i>Uvigerina</i>		OS-119052	10,300	30	
GGC43	180	6.2		<i>G. ruber</i>	OS-119355	10,150	35	
GGC43	220	3.2	<i>Uvigerina</i>		OS-109364	12200	55	
GGC43	220	3.9		<i>G. ruber</i>	OS-99079	11150	70	DWI
GGC43	280	3.6		<i>G. ruber</i>	OS-99080	12600	75	
GGC43	320	4.1		<i>G. ruber</i>	OS-99081	13800	60	
GGC43	350	2.9		<i>G. ruber</i>	OS-99082	15500	70	
CDH41	0.25	1.9	<i>Uvigerina</i>		OS-101662	9560	50	P>B
CDH41	0.25	4.1		<i>G. ruber</i>	OS-99072	9580	50	
CDH41	4.5	4.8	<i>Uvigerina</i>		OS-113941	11900	50	
CDH41	4.5	4.4		<i>G. ruber</i>	OS-113942	10500	45	
CDH41	10	3	<i>Uvigerina</i>		OS-102464	14900	70	
CDH41	10	2.9		<i>G. ruber</i>	OS-102469	12150	100	DWI
CDH41	15.25	2.2	<i>Uvigerina</i>		OS-102712	15300	65	
CDH41	15.25	2.1		<i>G. ruber</i>	OS-102711	12550	45	
CDH41	17.25	2.6	<i>Uvigerina</i>		OS-113943	14600	70	
CDH41	17.25	6.6		<i>G. ruber</i>	OS-113944	12750	55	
CDH41	20.25	2.5	<i>Uvigerina</i>		OS-102467	14150	70	
CDH41	20.25	2.8		<i>G. ruber</i>	OS-102466	11850	65	
CDH41	25.25	6.9	<i>Uvigerina</i>		OS-113945	15500	75	
CDH41	25.25	4.5		<i>G. ruber</i>	OS-113946	13800	55	
CDH41	30	1.6	<i>Uvigerina</i>		OS-101663	15400	95	
CDH41	30	3.6		<i>G. ruber</i>	OS-99073	14350	60	Bova et al., 2015
CDH41	35.25	6.7	<i>Uvigerina</i>		OS-113947	16500	80	
CDH41	35.25	5.2		<i>G. ruber</i>	OS-113948	13950	60	DWI
CDH41	40.25	2	<i>Uvigerina</i>		OS-102714	14950	60	

CDH41	40.25	2.4		<i>G. ruber</i>	OS-102713	12700	50	Bova et al., 2015
CDH41	47.25	2.5	<i>Uvigerina</i>		OS-113949	16700	80	
CDH41	47.25	4.5		<i>G. ruber</i>	OS-113950	14300	70	DWI
CDH41	55.25	2	<i>Uvigerina</i>		OS-102716	17700	90	
CDH41	55.25	2.2		<i>G. ruber</i>	OS-102715	14450	60	DWI
CDH41	70	2.1	<i>Uvigerina</i>		OS-101664	18850	150	
CDH41	70	5.5		<i>G. ruber</i>	OS-99074	15650	80	
CDH41	90	2.1	<i>Uvigerina</i>		OS-102718	21000	130	
CDH41	90	2.6		<i>G. ruber</i>	OS-102717	18150	90	
CDH41	100.25	7.9	<i>Uvigerina</i>		OS-119356	22,300	160	
CDH41	100.25	5.8		<i>G. ruber</i>	OS-119357	18,750	100	
CDH41	115.25	2	<i>Uvigerina</i>		OS-101665	21600	210	
CDH41	115.25	5.1		<i>G. ruber</i>	OS-99075	19700	140	
CDH41	135	3.4	<i>Uvigerina</i>		OS-102413	23100	220	
CDH41	135	2.5		<i>G. ruber</i>	OS-102719	20400	120	
CDH41	170	2.2	<i>Uvigerina</i>		OS-101667	23500	240	
CDH41	170	3.2		<i>G. ruber</i>	OS-99076	20600	120	
CDH41	215.25	2.2	<i>Uvigerina</i>		OS-118822	24,200	350	
CDH41	215.25	4.7		<i>G. ruber</i>	OS-119358	14200	60	Bova et al., 2015
CDH41	235.25	3.1	<i>Uvigerina</i>		OS-109457	25500	140	
CDH41	235.25	4.6		<i>G. ruber</i>	OS-109334	17950	70	Bova et al., 2015
CDH41	255.25	2.5	<i>Uvigerina</i>		OS-118823	29,800	710	
CDH41	255.25	3.5		<i>G. ruber</i>	OS-119359	24700	210	
CDH41	275.25	5.4	<i>Uvigerina</i>		OS-109287	28100	180	
CDH41	275.25	3.3		<i>G. ruber</i>	OS-109363	24000	120	Bova et al., 2015
CDH41	290.25	4.4	<i>Uvigerina</i>		OS-109362	27000	140	
CDH41	290.25	2.9		<i>G. ruber</i>	OS-109458	27200	190	Bova et al., 2015
CDH41	310.25	2.8	<i>Uvigerina</i>		OS-119362	28,100	320	
CDH41	310.25	4.5		<i>G. ruber</i>	OS-119360	26200	260	
CDH 26	96	5		<i>N. dutertrei</i>	OS-84131	3970	30	
CDH 26	171.5	6		<i>N. dutertrei</i>	OS-84109	5980	35	
CDH 26	203.5	3		<i>N. dutertrei</i>	OS-84125	6320	35	
CDH 26	235.5	5.4	<i>uvigerina sp.</i>		OS-103000	7620	40	
CDH 26	235.5	6		<i>N. dutertrei</i>	OS-84164	7070	35	
CDH 26	267.5	8		<i>N. dutertrei</i>	OS-84113	7440	35	
CDH 26	325	6		<i>N. dutertrei</i>	OS-84111	8550	40	
CDH 26	421	1.8	<i>uvigerina sp.</i>		OS-103151	9840	45	
CDH 26	421	6		<i>N. dutertrei</i>	OS-84110	9460	40	
CDH 26	478	8		<i>N. dutertrei</i>	OS-84159	10150	55	

CDH 26	574	1.9	bolivina sp.		OS-103153	11700	55
CDH 26	574	9		N. dutertrei	OS-84112	10750	45
CDH 26	663	2.7	uvigerina sp.		OS-103181	12400	40
CDH 26	663	8		N. dutertrei	OS-84161	11650	45
CDH 26	784.5	1.8	mixed benthic		OS-103150	13500	75
CDH 26	784.5	7		N. dutertrei	OS-84160	12350	50
CDH 26	848.5	9		N. dutertrei	OS-84165	12750	50
CDH 26	939.5	1.4	mixed benthic		OS-104319	14,200	90
CDH 26	939.5	9		N. dutertrei	OS-84156	13100	60
CDH 26	971.5	3.8		N. dutertrei	OS-101831	13050	55
CDH 26	1156.5	1.7	mixed benthic		OS-103154	14800	65
CDH 26	1156.5	4		N. dutertrei	OS-84128	13900	60
CDH 26	1299	6.3		N. dutertrei	OS-101830	14750	65
CDH 26	1400.5	4	uvigerina sp.		OS-103182	16000	60
CDH 26	1400.5	4		N. dutertrei	OS-84132	15150	65
CDH 26	1496.5	8.4		N. dutertrei	OS-101835	15550	75
CDH 26	1619.5	9		N. dutertrei	OS-84117	16250	65
CDH 26	1741	3.8	uvigerina sp.		OS-103183	18050	70
CDH 26	1741	7		N. dutertrei	OS-84114	17050	70
CDH 26	1895	2.2	uvigerina sp.		OS-103155	18950	100
CDH 26	1895	6		N. dutertrei	OS-84118	17700	70
CDH 26	2048	5.5	uvigerina sp.		OS-103001	19700	90
CDH 26	2048	9		N. dutertrei	OS-84158	18850	110
CDH 26	2169	6		N. dutertrei	OS-84119	20200	85
CDH 26	2265	3		N. dutertrei	OS-84129	19700	85
CDH 26	2386	2.5	mixed benthic		OS-103152	22300	160
CDH 26	2386	5		N. dutertrei	OS-84163	20900	80
CDH 26	2418	5.4		N. dutertrei	OS-101832	21500	150
CDH 26	2539	3		N. dutertrei	OS-84126	21900	95

*DWI = Deep water influence

APPENDIX D: Coccolith Relative Abundances

Core-top Calibration Dataset:

Core-top ID	N ratio	Latitude	Longitude	Mean Annual Thermocline Depth (m)
ME0005A-38MC	0.494339623	7.3167	-84.113	36.25
ME0005A-14MC	0.355978261	5.8465	-86.449	42.5
046Tri-RC1111	0.512195122	-4.8333	-86.825	41.25
ME0005A-15MC	0.4	4.6137	-86.704	45
ME0005A-20MC	0.431818182	3.2123	-86.486	40
EQP-3a	0.409937888	-0.0448	-105.426	40
ME0005A-29MC	0.663082437	-0.5134	-81.995	27.5
KNR195-05-04-09C-MC	0.474226804	-0.826	-87.90817	27.5
ME0005A-25MC	0.568106312	-1.8534	-82.787	23.75
KNR195-05-13-33D-MC	0.407407407	-3.22358	-82.91417	28.75
KNR195-05-14-34C-MC	0.474666667	-3.59753	-83.96317	31.25
KNR195-5 22MC	0.684745763	-3.8504	-81.25	13.75
KNR195-05-08-18A-MC	0.667808219	-3.98662	-81.3099	16.25
043Tri-RC1111	0.30125523	-11.3333	-106.9667	75
045Tri-RC1111	0.284482759	-11.6	-95.6333	78.75
044Tri-RC1111	0.279411765	-13.3167	-100.95	98.75
KNR195-05 MC 12	0.636666667	-3.71057	-81.10383	13.75
KNR 195-5 25MC	0.75	-3.58253	-81.1645	13.75

Galapagos Core-tops

KNR195-05-17-42A-MC	0.288951841	-1.25302	-89.68555	25
KNR195-05-02-04B-MC	0.325342466	-0.21613	-89.50422	26.25
KNR195-05-05-11A-MC	0.37458194	-1.53258	-86.78517	25

Down-core Thermocline Depth Records:

ME0005A-15MC/17JC:

Core ID	Core Depth (cm)	Age (yrs BP, Marine13)	N ratio	Thermocline Depth (m)
ME5-15MC	5.25	8099	0.492	28.0
ME5-15MC	5.25	8099	0.519	25.9
ME5-15MC	8.25	9573	0.514	26.3
ME5-15MC	10.25	10556	0.501	27.2
ME5-17JC	8.25	10687	0.485	28.6
ME5-17JC	10.25	11774	0.480	29.1

ME5-15MC	15.25	12522	0.550	24.0
ME5-17JC	12.25	12849	0.519	25.9
ME5-17JC	14.25	13913	0.536	24.9
ME5-17JC	16.25	14966	0.539	24.7
ME5-17JC	18.25	16007	0.588	22.3
ME5-17JC	20.25	17038	0.667	19.8
ME5-17JC	22.25	18057	0.618	21.2
ME5-17JC	24.25	19064	0.689	19.3
ME5-17JC	26.25	20061	0.444	33.1
ME5-17JC	26.25	20061	0.571	23.0
ME5-17JC	28.25	21046	0.586	22.4
ME5-17JC	30.25	22020	0.549	24.1
ME5-17JC	32.25	22982	0.592	22.1
ME5-17JC	34.25	23934	0.472	29.9
ME5-17JC	36.25	24874	0.434	34.4
ME5-17JC	40.25	26720	0.456	31.6
ME5-17JC	42.25	27627	0.456	31.6
ME5-17JC	44.25	28522	0.502	27.2
ME5-17JC	46.25	29406	0.458	31.4
ME5-17JC	48.25	30278	0.505	26.9
ME5-17JC	52.25	31989	0.531	25.2

GGC 43/CDH 41:

Core ID	Core Depth (cm)	Age (yrs BP, Marine13)	<i>N ratio</i>	Thermocline Depth (m)
GGC 43	2.25	1055	0.411	41.3
GGC 43	26.25	3124	0.353	54.2
GGC 43	42.25	4313	0.461	33.6
GGC 43	60.25	5493	0.383	46.9
GGC 43	74.25	6308	0.434	37.4
GGC 43	90.25	7143	0.414	40.7
GGC 43	120.25	8472	0.431	37.8
GGC 43	138.25	9147	0.359	52.7
GGC 43	160.250	9875	0.435	37.3
GGC 43	202.25	11078	0.418	39.9
GGC 43	218.25	11514	0.502	28.8
GGC 43	234.25	11962	0.540	25.2
GGC 43	280.25	13501	0.582	22.0
GGC 43	298.25	14297	0.435	37.2

GGC 43	325.25	15833	0.436	37.0
GGC 43	338.25	16763	0.545	24.7
GGC 43	340.25	16920	0.549	24.4
GGC 43	346.25	17411	0.556	23.9
GGC 43	360.25	18704	0.543	25.0
CDH 41	82.25	20151	0.475	31.8
CDH 41	98.25	21162	0.506	28.4
CDH 41	122.25	22403	0.488	30.2
CDH 41	146.25	23442	0.596	21.1
CDH 41	170.25	24389	0.470	32.3
CDH 41	194.25	25315	0.582	22.0
CDH 41	238.25	27054	0.611	20.1
CDH 41	270.25	28290	0.645	18.3
CDH 41	286.25	28842	0.678	16.7

VNTR01-13GC:

Core ID	Core Depth (cm)	Adjusted Core Depth (cm)	Age (yrs BP, Marine13)	<i>N ratio</i>	Thermocline Depth (m)
VNT 13GC	2.25	2.25	764	0.323	65.3
VNT 13GC	8.25	8.25	4682	0.290	86.6
VNT 13GC	12.25	12.25	6664	0.309	73.0
VNT 13GC	17.25	17.25	8615	0.406	39.1
VNT 13GC	20.25	20.25	9570	0.404	39.5
VNT 13GC	23.25	23.25	10401	0.442	33.3
VNT 13GC	28.25	28.25	11589	0.363	49.5
VNT 13GC	32.25	32.25	12419	0.490	28.2
VNT 13GC	36.35	36.35	13211	0.612	21.4
VNT 13GC	44.25	44.25	14690	0.566	23.2
VNT 13GC	48.25	48.25	15458	0.628	20.9
VNT 13GC	52.25	52.25	16255	0.509	26.6
VNT 13GC	92.25	56.25	17086	0.570	23.1
VNT 13GC	96.25	60.25	17951	0.522	25.7
VNT 13GC	100.25	64.25	18843	0.462	30.9
VNT 13GC	110.25	74.25	21133	0.491	28.0
VNT 13GC	122.25	86.25	23738	0.433	34.6
VNT 13GC	124.25	88.25	24130	0.399	40.5
VNT 13GC	130.25	94.25	25197	0.384	43.9
VNT 13GC	132.25	96.25	25512	0.462	30.9
VNT 13GC	134.25	98.25	25804	0.399	40.6

VNT 13GC	138.25	102.25	26316	0.443	33.2
VNT 13GC	148.25	112.25	27165	0.556	23.7
VNT 13GC	150.25	114.25	27264	0.475	29.5
VNT 13GC	154.25	118.25	27403	0.533	25.0
VNT 13GC	158.25	122.25	27479	0.580	22.6
VNT 13GC	166.25	130.25	27556	0.609	21.5
VNT 13GC	170.25	134.25	27638	0.507	26.8