Climate-related response of dust flux to the central equatorial Pacific over the past 150 kyr

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9 **Abstract**

10 High resolution paleoclimate records from low latitudes are critical for understanding the
11 role of the tropics in transmitting and generating feedbacks for high-latitude climate change on
12 glacial-interglacial and millennial timescales. Here we present three new records of $^{230}\text{Th}_{\text{xs},0}$ -
13 normalized ^{232}Th -derived dust fluxes from the central equatorial Pacific spanning the last 150
14 kyr at millennial-resolution. All three dust flux records share the “sawtooth” pattern
15 characteristic of glacial-interglacial cycles in ice volume, confirming a coherent response to
16 global climate forcing on long timescales. These records permit a detailed examination of
17 millennial variability in tropical dust fluxes related to abrupt perturbations in oceanic and
18 atmospheric circulation. Increases in dust flux in association with at least six of the longest
19 Greenland stadials provide evidence that abrupt, high-latitude climate oscillations influenced the
20 atmospheric aerosol load in the equatorial Pacific, with implications for both direct and indirect
21 effects on the tropical energy balance. Our latitudinal transect of cores captures shifts in the
22 position of the Intertropical Convergence Zone (ITCZ) in response to variations in the
23 interhemispheric thermal gradient associated with cooling in Greenland and bipolar seesaw
24 warming in Antarctica. These observations demonstrate that changes in the energy and
25 hydrologic balance of the tropics were repeated features of the penultimate deglaciation, last
26 glacial inception and last glacial cycle, and highlight the role of the tropical atmosphere as a
27 dynamic and responsive component of Earth’s climate system.

28 **Keywords**

29 Dust, Abrupt climate change, Stadial events, Intertropical Convergence Zone, Equatorial Pacific

1. Introduction

Windblown mineral dust plays a critical role in the climate system through its alteration of radiative forcing, surface albedo, significance as cloud and ice condensation nuclei, and ability to enhance biological sequestration of carbon (review in Mahowald et al., 2014). Long, high-resolution records of mineral dust flux exist from sites in Antarctica (Lambert et al., 2012) and Greenland (Ruth et al., 2007), but no equivalent reconstructions exist from the tropics where the atmospheric load of dust can be especially important due to high incoming solar radiation and large convective potential. Previous work has reconstructed dust fluxes in the tropics at low resolution (Anderson et al., 2006; McGee et al., 2007), and has been critical in establishing that global dust fluxes demonstrate coherent cyclicity over at least the last 500 kyr (Winckler et al., 2008). Despite these advances, the temporal and spatial resolution of existing tropical dust flux records has limited our ability to draw conclusions about low latitude climate responses and feedbacks. For example, the identification of millennial stadial events in equatorial dust fluxes has been challenging due to insufficient data resolution. Abrupt dust flux changes that would be smoothed out by bioturbation in low sedimentation rate environments, or hidden in noisy proxy reconstructions are important for interpreting the far-field implications of high latitude climate shifts with significant consequences for tropical radiative balance. Additionally, although geographically proximal dust flux reconstructions have the potential to inform our understanding of spatially variable dust removal processes the temporal span of such records is limited (McGee et al., 2007). Multiple archives yielding proxy data with high temporal resolution and a low signal to noise ratio have the potential to reveal information about regional climate variability. The value of this type of approach was recently demonstrated using dust flux records to identify major shifts in tropical hydrology at the termination of the penultimate glacial period (Jacobel et

al., 2016). Reconstructing records of tropical dust fluxes over glacial cycles at sub-millennial scale resolution is important as we try to understand linkages between high and low latitude climate and the mechanisms of climate forcing and feedbacks.

In tropical ocean basins, far from continental sources, the most important removal mechanism for dust is wet deposition by the Intertropical Convergence Zone (ITCZ), which efficiently scavenges particles from the atmospheric column (McGee et al., 2007; Schlosser et al., 2014) and acts as a barrier to interhemispheric dust transport (Xie and Marcantonio, 2012; Pichat et al., 2014). The mean position of the ITCZ is often identified via satellite imagery as the latitude at which precipitation, and consequently wet-deposition of aerosols (Schlosser et al., 2014), is at an annual maximum. The ITCZ forms where northern and southern hemisphere trade winds converge and convection occurs. On long timescales, the mean position of the ITCZ is responsive to the latitude of Earth's thermal equator (Schlosser et al., 2014) with the ITCZ shifting towards the differentially warming hemisphere. This behavior has been demonstrated in model simulations and paleoclimate reconstructions for glacial-interglacial climate shifts (Chiang et al., 2003; Arbuszewski et al., 2013), abrupt stadial events (Wang et al., 2004; Chiang and Bitz, 2005; Jacobel et al., 2016), and even changes during the historical period arising from hemispherically asymmetric aerosol loading (Ridley et al., 2015). This interplay of forces influencing ITCZ position adds complexity that makes a simple determination of ITCZ sensitivity to thermal perturbations difficult, hindering our predictive capabilities.

Addressing questions about past ITCZ shifts and their causes is important not only in establishing the range of responses to asymmetric climate forcing, but also because perturbations of the ITCZ may act to alter global atmospheric circulation patterns. Modeling work has shown that changes in ITCZ latitude influence the position of the Southern Hemisphere Westerlies (Lee

et al., 2011), a key component of the Southern Ocean deglacial upwelling hypothesis (Anderson et al., 2009). Recent work has indicated that a southward shift of the Pacific ITCZ at least 4.5 degrees south of its present day latitude accompanied the deglacial stadial event that punctuated the penultimate deglaciation (~150-120 ka) (Jacobel et al., 2016). This shift of the ITCZ may have driven significant changes in global atmospheric circulation patterns, contributing to the conditions required for deglaciation. In sum, existing evidence suggests the position of the ITCZ is a key indicator of and intermediary player in global climate change, making long, high spatiotemporal resolution records of its behavior essential in developing a comprehensive picture of past climate.

Here we present high-resolution records from a latitudinal transect of three sediment core sites in the central Equatorial Pacific to constrain ITCZ movement and position via the quantification of dust fluxes from the penultimate deglaciation, through the last glacial inception, last glacial maximum and last deglaciation. Our records are the first to constrain equatorial Pacific dust fluxes at sub-millennial resolution over this entire time period, allowing us to examine the sensitivity of both tropical dust fluxes and the ITCZ to a range of perturbations, occurring over a variety of timescales. We focus our discussion on the last glacial inception and the last glacial period ~ 120-20 ka, as we have previously considered the penultimate deglaciation at high resolution (Jacobel et al., 2016) and a detailed comparison of the last two deglaciations (Terminations I and II) is forthcoming (Jacobel et al., in prep.).

2. Study Area, Sediment Cores and Age Models

Three cores were selected for this study, forming a north-south transect in the central equatorial Pacific (cruise ML1208) (Figure 1). All cores were taken near the Line Islands archipelago, a north-south trending chain of atolls. Core ML1208-37BB (hereafter 37BB) was

recovered from 2,798 m of water at 7.04°N, 161.63°W, core ML1208-31BB (31BB) from 2,857 m of water at 4.68°N, 160.05°W and core ML1208-17PC (17PC) at 0.48°N, 156.45°W from a water depth of 2,926 m. Core sediments are composed primarily of foraminifer sands with correspondingly high calcium carbonate content. Site 37BB is located near the maxima in mean annual precipitation associated with the modern ITCZ (~7°N) (at 160°W), site 31BB is just south of both the boreal winter and mean annual precipitation maxima, and site 17PC is consistently south of the seasonally shifting ITCZ (Figure 1).

Age models for the cores (Supplementary Figure 1) are based on *G. ruber* (planktonic) foraminifera $\delta^{18}\text{O}$ isotope stratigraphies and radiocarbon dates (Lynch-Stieglitz et al., 2015), a portion of which were newly obtained for this study (Supplementary Note 1 and Table 1). Planktonic isotope stratigraphies were used to constrain chronologies beyond the range of radiocarbon via tuning to the LR04 benthic stack (Lisiecki and Raymo, 2005) using MonteXCM (Lisiecki and Raymo, 2005; Hoffman et al., 2015), making these chronologies distinct from those previously published by Lynch-Stieglitz et al., 2015. MonteXCM is a modified Monte-Carlo-enabled cross-correlation maximization scheme coupled with a random walk algorithm to assign age-depth relationships at control points based on the statistical match between the input data (planktonic $\delta^{18}\text{O}$) and a target chronology (the LR04 stack) within the analytical uncertainties of both data sets. Age control points from MonteXCM were passed into Bchron's 'Bchronology' function (Haslett and Parnell, 2008) to better constrain the age uncertainty associated with each data point using the Compound Poisson-Gamma chronology model available in Bchron. Resulting age model uncertainties average ± 4.1 , 2.8 and 4.6 ka for 37BB, 31BB and 17PC respectively (2 sigma). Because each age model is developed independently from the others, small offsets between the timing of dust flux events exist between the records.

However, we have no reason to suspect that these differences represent significant lead or lag relationships between the three sites and attribute these apparent differences to the uncertainties inherent in our age models. Unless otherwise noted, ice core data used for data comparison are presented on the AICC2012 age model (Veres et al., 2013). General comments on the significance of accumulation rate differences and bioturbation in the ML1208 cores can be found in Supplementary Note 2 while interval-specific details are in Section 6.4.2, where the signatures of millennial events are discussed and interpreted.

3. Proxies and Application

3.1 ^{232}Th as an aeolian dust proxy

In this study we employ $^{230}\text{Th}_{\text{xs},0}$ -normalized ^{232}Th fluxes as a proxy for dust accumulation at our sites. Lithogenic sediments have been shown to contain ^{232}Th at average concentrations of 10.7 ppm varying 1 ppm or less over a wide range of provenance (McGee et al., 2007) including those areas likely to be sources for our sites (Supplementary Note 2). The quantification of $^{230}\text{Th}_{\text{xs},0}$ -normalized ^{232}Th fluxes as a terrestrial dust proxy is also supported by work that demonstrates a tight correlation between ^{232}Th and terrigenous ^4He (Winckler et al., 2005; McGee et al., 2016).

3.2 $^{230}\text{Th}_{\text{xs},0}$ normalization

Sediment mass accumulation rates and dust fluxes were evaluated by normalization to $^{230}\text{Th}_{\text{xs},0}$ (Anderson et al., 2006) (Supplementary Note 3 and references therein). The fundamentals underlying the use of ^{230}Th -normalization to account for dilution and post-depositional sediment changes have been the subject of much research and debate, with recent work suggesting that $^{230}\text{Th}_{\text{xs},0}$ may be problematic for reconstructions of coarse fraction MAR in

areas where focusing factor corrections are large (Lyle et al., 2014). Our study avoids the potential pitfalls identified by these authors because we use ^{230}Th to normalize ^{232}Th concentrations and both isotopes reside predominately in the fine fraction. This reasoning is in accord with results suggesting that Al fluxes (another fine fraction constituent) from sediment traps are in agreement with ^{230}Th -derived focusing factors (Lyle et al., 2014).

3.3 $^{232}\text{Th}/^{230}\text{Th}_{\text{xs},0}$ (Dust flux) as a proxy for ITCZ position

Open ocean $^{230}\text{Th}_{\text{xs},0}$ -normalized ^{232}Th fluxes record the flux of aeolian dust. Mineral dust is preferentially removed from the atmosphere in areas of high precipitation, like those underlying the ITCZ. Observations of ocean chemistry in the tropical Atlantic Ocean show a linear correlation between ITCZ precipitation and dissolved iron from mineral dust (Schlosser et al., 2014). Existing paleorecords confirm that sites within the ITCZ are characterized by latitudinal gradients in dust flux (McGee et al., 2007) and provenance (Xie and Marcantonio, 2012; Reimi and Marcantonio, 2016). Thus, we can use the scavenging of dust by the ITCZ to determine its position, taking into consideration that dust at the northern edge of the ITCZ will be primarily sourced from the Northern Hemisphere (NH) and that which is deposited at its southern edge will be predominately sourced from the Southern Hemisphere (SH). Here, we compare our equatorial dust flux records to independent records of dust flux from Greenland, the Southern Ocean, and Antarctica. These records, combined with our latitudinal transect approach, allow us to differentiate between changes in atmospheric dust abundance and changes in ITCZ position.

4. Methods

A total of 604 samples from cores 37BB, 31BB and 17PC (109, 286 and 209 respectively) were analyzed for uranium and thorium by isotope dilution and inductively coupled plasma mass

spectrometry on an Element 2 ICP-MS at the Lamont-Doherty Earth Observatory of Columbia University. For each sample 100 to 200 mg of sediment was spiked with ^{236}U and ^{229}Th prior to dissolution and digestion using HNO_3 , HClO_4 and HF (Fleisher and Anderson, 2003). Anion exchange column chemistry was used to isolate the U and Th fractions from the bulk digestate (Kraus et al., 1956). Samples were run with an internal sediment standard to determine measurement reproducibility ($n=29$), yielding relative standard deviations of 2.7% for ^{238}U , 2.3% for ^{230}Th and 3.7% for ^{232}Th . Background contamination was evaluated in each run via blanks that were spiked and carried through column chemistry, but contained no sediment. Results show ^{238}U blanks (0.3 ng), ^{230}Th blanks (0.09 pg) and ^{232}Th blanks (0.8 ng), values that are smaller than 1.2% of even the lowest sample values. Errors reported in the figures represent 2σ uncertainty specific to each sample, including the propagated error due to uncertainty in counting statistics, mass bias corrections, counting gain corrections, spike measurements and uncertainty in the fraction of sediment that is lithogenic in origin (Bradtmiller et al., 2010).

5. Results

5.1 Mass accumulation rates

Data used to derive ^{232}Th -based dust fluxes are presented in Figure 2 for each of our three new sediment core records. Variations in $^{230}\text{Th}_{\text{xs},0}$ -based mass flux are displayed in panel 2 (a) and account for changes in the rain of particulate matter from the surface after correcting for sediment focusing and winnowing (Supplementary Note 4). The data display both millennial variability and long-term changes in sedimentation, but mass fluxes vary similarly between the three core sites, with fluxes highest at the equator and decreasing with increasing latitude. This pattern is as expected given enhanced organic and calcareous productivity associated with equatorial upwelling.

The importance of flux-normalizing concentration data is readily illustrated by a comparison of Figure 2 (b) with (c), demonstrating that mass flux normalization removes sedimentary effects which might otherwise create the false impression of differences between the magnitude of dust deposition between cores and within cores. The value and significance of the mass flux correction is especially evident in 37BB in which ^{232}Th concentrations approach values twice the magnitude reconstructed in the other two cores. Higher concentrations of ^{232}Th at site 37BB are due primarily to the lower rain of biogenic particles that dilute the dust concentration there less compared to other sites. Normalization to $^{230}\text{Th}_{\text{xs},0}$ corrects the ^{232}Th concentration for variable dilution by biogenic phases, which contain negligible amounts of ^{232}Th .

5.2 Dust fluxes

Reconstructed dust fluxes show the largest amplitude changes on 100 kyr timescales in all three sediment cores, with smaller but well-resolved oscillations on millennial time scales occurring during both glacial and interglacial periods (Figure 3). The highest dust fluxes are observed during glacial MIS 2, 4 and 6, (29-14, 71-57 and 191-130 ka respectively) and the lowest fluxes are recorded during interglacial MIS 1 and 5 (14-0 and 130-71 ka respectively). Dust fluxes increase through the glacial inception (~115 ka), continue to increase slowly through the remainder of MIS 5, rise sharply during MIS 4 and then experience a termination-like decrease from MIS 4 into MIS 3. Indeed, early MIS 3 dust fluxes are at less than full glacial levels, although still greater than at the end of MIS 5. Dust fluxes rise somewhat into the MIS 2 and then decline dramatically during glacial Termination (T) I as they do during TII at the end of MIS 6. All three cores indicate that MIS 4 was as dusty or dustier than MIS 2, with some variability in the amount of dust present at MIS 4 relative to MIS 6.

Superimposed on this orbital variability are millennial changes in dust flux including (1) an increase in dust flux at the time of Heinrich Stadial (HS) 11 (~136-129 ka) (most notably in 17PC and 31BB), (2) small but coherent and well-resolved oscillations in dust flux at 31BB from ~109.3-103.7, 114.1-109.3 and 121.0-117.1 ka, and (3) an increase in dust flux between approximately 93 and 81 ka in cores 31BB and 17PC with a “late” intensification in dust fluxes at 37BB from ~88 to 74 ka. Some of these events are visible only at the highest sedimentation rate site (31BB) possibly due to bioturbation, lower sediment accumulation rates, or to the barrier effects of the ITCZ on cross convergence zone dust transport. These alternatives will be evaluated in detail in section 6.4.2.

Because our analytical uncertainty is smaller than the differences between our three records at several intervals over the 150 kyr record, we are also able to examine spatial gradients in dust flux across our section (Fig 4c). High HS11 dust fluxes in the equatorial core 17PC have previously been discussed (Jacobel et al., 2016), so here we focus on the data from 120 ka to the last glacial maximum. Most notably, during MIS 4 significantly higher dust fluxes are observed at 37BB, the northernmost and lowest sedimentation rate site, relative to those at 31BB or 17PC. An average of the three highest MIS 4 dust flux values at each site shows that fluxes in 37BB are 20% higher than 31BB and 16% higher than at 17PC.

6. Discussion

6.1 Coherent glacial-interglacial variability

Previous studies have noted a strong covariance on long timescales between eastern equatorial Pacific dust fluxes, Antarctic dust fluxes and global ice volume (as reconstructed by the LR04 benthic stack), using this observation to suggest that dust generation processes in each hemisphere respond similarly to global climate changes over at least the last 500 kyr (Winckler

et al., 2008). Here, a comparison of our new data to highly resolved records of dust flux from Greenland (Ruth et al., 2007) and Antarctica (Lambert et al., 2012) (Figure 4), as well as the Southern Ocean (Anderson et al., 2014) (Figure 6e), are consistent with this observation. Broadly, the glacial-interglacial pattern of dust fluxes shows low interglacial dust fluxes with glacial dust fluxes ~3 times higher. Although this correlation is striking, it does not directly illuminate the element(s) of global climate variability driving changes in the dust flux. In a comprehensive review of the drivers of glacial dustiness, McGee and coauthors suggest that the most important factor may be increased wind gustiness which is dependent upon meridional thermal gradients and increased high and mid latitude seasonality (McGee et al., 2010). Changes in wind gust frequency and maximum speed may thus be driven by the growth of ice at the poles or by orbital variability in insolation gradients (and the seasonality of maximum contrast), inducing large scale changes in atmospheric circulation that translate on a regional scale to enhanced dust entrainment and transport. In this context it is interesting to observe that equatorial dust fluxes more closely track “sawtooth” changes in global ice volume than do high latitude dust flux records from the poles, which show similar baseline dust fluxes during MIS 5 and MIS 3 (Figure 4 b and d). Enhancement of westerly winds during glacial periods has previously been demonstrated in modeling experiments that attributed the strengthening primarily to the growth of ice sheets rather than a reduction in CO₂ (summary in Chiang and Friedman, 2012).

6.2 Dusty Details and Millennial Events

Whereas previous studies have noted the strong first-order correlation between high latitude and tropical dust flux records on ~100 kyr timescales (Winckler et al., 2008), the increased temporal resolution of our new records allows for a more detailed evaluation and reveals shorter-

lived features in the equatorial Pacific records that are correlative with events also observed at high-latitudes. Recent work, using the same sediment cores and techniques employed here, reconstructed a previously unknown millennial dust flux in association with the penultimate deglaciation and Heinrich Stadial 11 (Jacobel et al., 2016) and previous work in the far eastern equatorial Pacific identified an increase in dust in concert with Heinrich Stadial 1 (Kienast et al., 2013). Here we identify the signature of other stadial events that influenced equatorial Pacific dust fluxes over the last 150 kyr. In this section, we emphasize observations made using our highest resolution record, 31BB, focusing first on the identification of millennial increases in dust flux and then discussing the relationship between millennial stadial events and ITCZ movement in Section 6.4.2.

Stadial events have several different naming conventions depending on their origin and the paleoclimate archive in which they are identified. For example, intervals of depleted $\delta^{18}\text{O}$ corresponding to cold temperatures reconstructed from Greenland ice cores are called Greenland Stadials (GS) (Rasmussen et al., 2014). These GS events are frequently accompanied by increases in the flux of dust to ice core sites and sometimes include Heinrich Events, recognized as intervals of high ice rafted debris (IRD) in marine sediment cores, or C-Events characterized by intervals of cold SST in marine sediment records (McManus et al., 1994). The term Heinrich Stadial (HS) is often used to describe a cold interval that includes a Heinrich Event but is characterized by a proxy other than IRD. Previous work has established that nearly all intervals of widespread cooling in the NH correspond with intervals of warmth in Antarctica, a phenomenon known as the bipolar seesaw (review in Landais et al., 2015). The exception to this rule is event GS26 (Capron et al., 2012), which occurs during the glacial inception when temperatures are decreasing globally. As a consequence of paired NH warming and SH cooling,

most Greenland stadial events are associated with perturbations of the interhemispheric thermal gradient and are therefore thought to have resulted in shifts of the ITCZ (Landais et al., 2015).

6.2.1 Greenland Stadials 24, 25 and 26

During the last glacial inception beginning at ~123 ka, three stadial events occurred which are recognized in Greenland ice as GS 24, 25 and 26 (~105.4-104.5, 110.6-108.3 and 119.1-115.4 ka on the GICC05 age model of (Seierstad et al., 2014)) (~103.2-102.2, 107.9-105.6 and 116.5-112.3 ka on the AICC2012 age model (Veres et al., 2013) (Figure 5). These events have been identified both as intervals of cold temperatures in Greenland and cool SSTs in the North Atlantic (McManus et al., 1994; Oppo et al., 2006) and subtropical North Atlantic (Lehman et al., 2002; Pailler and Bard, 2002). Unfortunately, dust flux records from Greenland only extend to 108 ka and thus capture only GS 24 and its associated dust flux increase. Within the uncertainty of our age model and the two different age models for the ice core, the three GS events correspond with dust flux increases in core 31BB from ~109.3-103.7, 114.1-109.3 and 121.0-117.1 ka (Figure 5). The three abrupt GS events appear somewhat more gradual in our records, likely due to the influence of bioturbation. Indeed, in a modeling exercise simulating bioturbation of an abrupt input function under constraints similar to those in 31BB, the synthetic dust flux record clearly mirrors our reconstructed dust fluxes (Supplementary Note 6) (Supplementary Figure 3).

The identification of GS 24, 25 and 26 in the tropics is an intriguing result because it provides evidence that even millennial perturbations in ocean circulation had an impact on tropical climate via atmospheric dust abundance and associated climate variables. This conclusion complements data from the eastern equatorial Pacific suggesting ITCZ moisture shifts occurred in concert with North Atlantic stadials over the last 90 ka (Leduc et al., 2007).

We suggest that abrupt climate perturbations previously identified in paleoclimatic records from Greenland and the North Atlantic, including those during MIS 5, may have had a more significant effect on Earth's radiative balance and other climate variables than previously appreciated.

6.2.2 Greenland Stadial 22

From 87.2 to 84.0 ka the Northern Hemisphere experienced a stadial event known in various climate archives as GS22, HS8 or C21 (Figure 6) (Lehman et al., 2002; Rasmussen et al., 2003; 2014). Greenland ice records indicate that this was a time of cold temperatures and increased dust fluxes (North Greenland Ice Core Project members, 2004; Ruth et al., 2007), North Atlantic sediments show cooling and the deposition of ice-rafted debris (McManus et al., 1994), and speleothem data from Sanbao, China are interpreted as indicating a weakened East Asian summer monsoon over the same interval (Wang et al., 2008). The abrupt dust flux increase in Greenland during GS22 is in contrast with data from East Antarctica (Lambert et al., 2012) and the Southern Ocean (Anderson et al., 2014), which shows GS22 to be an interval of declining dust fluxes, preceded by an Antarctic dust increase (Figure 6). Because these two ice core records have had their age models synchronized using methane matching (Veres et al., 2013), we can be confident that the dissimilar dust fluxes observed in opposite hemispheres are indeed two separate events.

The interval between 95 and 85 ka is exceptionally well resolved in core 31BB with 34 data points yielding an average resolution of 350 years (not including replicates) (Figure 6d). Reconstructed dust fluxes exhibit a double peak with local maxima at ~90.7 and 87.0 ka. The more recent of these two dust peaks is almost certainly correlative with GS22/HS8/C21, but attribution for the earlier peak is more difficult. For the older dust flux maxima to be correlative

326 with the SH dust increased observed in Antarctica, sedimentation rates would need to approach
327 12 cm ka^{-1} . Given the pelagic setting of this core and the observation that sedimentation rates at
328 the site are more typically in the $3\text{-}5 \text{ cm ka}^{-1}$ range, we find this scenario unlikely. Alternatively,
329 the early dust flux peak in 31BB may be correlative with GS23/C22. However, in Greenland
330 GS23/C22 is characterized by only a small oxygen isotope anomaly (North Greenland Ice Core
331 Project members, 2004) and has no observable dust flux signature, in contrast with GS22/C21
332 which is identified by a larger isotope anomaly and significantly enhanced dust accumulation. In
333 the subpolar North Atlantic C22 is similarly muted, being characterized by only a small decline
334 in SST and negligible IRD deposition (McManus et al., 1994; Oppo et al., 2006). In contrast,
335 SST underwent a brief, sharp decline in the subtropical Atlantic (Lehman et al., 2002) and
336 Chinese speleothems record a decline in East Asian monsoon precipitation during GS23/C22
337 (Wang et al., 2008). The mismatch between records from the high and subtropical latitudes thus
338 leaves open the possibility that GS23 was a significant event in NH climate that had particular
339 implications for the tropics. If this is the case then it would appear that dust associated with
340 GS23 did not reach Greenland, possibly due to anomalous atmospheric circulation. For site
341 31BB to approximately match the relative timing and duration of GS23 and GS22, as recorded at
342 Sanbao Cave, would require a change in sedimentation rate ($\sim 2 \text{ cm ka}^{-1}$ increase) precisely at the
343 trough between the two dust flux peaks. We find this to be a more likely scenario than the 12 cm
344 ka^{-1} rate required for the peak to be of Southern Hemisphere provenance. Constraining the origin
345 of the pre-GS22 dust peak is an important goal because 1) an abrupt shift of the hemispheric
346 origin of dust deposited at site 31BB might indicate an abrupt, if small, shift of the ITCZ and 2)
347 if the early dust flux peak we have identified is associated with GS23, it points to changes in the

atmospheric pathway associated with the transport of Asian dust to Greenland, which may have been a unique feature of GS23.

Site 37BB shows a different pattern from the two more southerly cores during GS22 and GS23 and immediately afterwards (Figure 3). We suggest that the reason for this difference is an age model problem that makes the age of the increase in dust flux observed at 37BB between ~88 and 74 ka too young. The calculated one sigma age model uncertainty for the onset of this event is 88 ± 2.5 ka, which means that this event is probably coincident in the two cores. Site 37BB does demonstrate a much longer event duration relative to 31BB which cannot be attributed solely to age model differences and instead likely reflects the smoothing effects of bioturbation due to the low accumulation rate of 1.2 cm ka^{-1} in 37BB during this interval.

6.2.3 Marine Isotope Stage 4

Our tropical dust flux records show that MIS 4 was at least as dusty as MIS 2, and our lowest accumulation rate core (37BB at 7.0°N) suggests that MIS4 was dustier than MIS 2 in agreement with the record from Greenland (excepting the two brief dust peaks associated with Heinrich Stadials 1 and 2). This is an interesting observation given globally lower ice volume, higher CO_2 concentrations and warmer temperatures during MIS 4 relative to MIS 2 (Petit et al., 1999). The apparent disconnect between observations of these climatic parameters thought to exert control on “dustiness” (summary in McGee et al., 2010) and the magnitude of dust flux observed at MIS 4 relative to MIS 2 leads to several hypotheses: 1) dust sources were more productive for other reasons (Rea, 1994), 2) uplift and transport of dust was more efficient (McGee et al., 2010), 3) additional sources of dust contributed to tropical and NH fluxes, or 4) depositional processes were stronger. Without provenance data for the dust at the ML1208 sites it is difficult to assess hypothesis 3, but a record of dust flux from the North Pacific (Serno, pers. comm., submitted)

also shows MIS 4 dust fluxes to be greater than during MIS 2, in support of the idea that increases in dust availability, entrainment and transport from Asian sources (hypotheses 1 and 2) may have played a significant role at latitudes north of, or within, the seasonally shifting ITCZ. This interpretation is strengthened by data indicating that during MIS 4 the median diameter of wind-blown quartz grains in China's Luochuan Loess increased (Porter and Zhisheng, 1995), and the flux of aeolian quartz to Japan's Lake Biwa was large (Figure 7c), coincident with a decrease in the flux of fluvial quartz to the same lake (Xiao et al., 1999). The authors of these records interpret the results to indicate that MIS 4 winter monsoon winds were stronger (Porter and Zhisheng, 1995; Xiao et al., 1999), and that the summer monsoon was weaker, leading to drier conditions (Xiao et al., 1999). Additional support for strengthened NH westerly winds comes from the Iberian margin where a sediment core taken from within the modern wind belt shows sea surface temperature cooling during MIS 4 to temperatures 3-4 degrees cooler than at the LGM (Pailler and Bard, 2002) (Figure 7b). Pailler and co-authors attribute this cooling to the combined influence of cold North Atlantic waters advected to the site during HS6 (63.55-59.23 ka) (North Greenland Ice Core Project members, 2004) and strengthened wind-driven upwelling.

Data also exist in support of strengthened MIS 4 southeast trade winds in the SH. For example grain size records from South Africa show enhanced coarse fraction delivery during MIS 4 above average MIS 2 values (Stuut et al., 2002). The SH may also have had active dust source areas that could have contributed dust to the central equatorial Pacific during its MIS 4 glacial maxima, but data are currently lacking to evaluate this hypothesis. In sum, more arid summers in Asia and globally strengthened winds during MIS 4 may have led to increased deflation and transport of dust, resulting in the larger dust fluxes observed in Greenland and at our equatorial sites.

The fourth mechanism for increased MIS 4 dust fluxes at the ML1208 sites is stronger tropical depositional processes. In light of the large increase in North Pacific dust fluxes coincident with those in the tropics (Serno, pers. comm., submitted), we suggest if depositional processes were enhanced they were not the primary factor contributing to higher dust fluxes. Although our records cannot be interpreted simply as indicative of the amount of precipitation or wet-scavenging occurring over our three sites, we do suspect hydrological changes may have played a role in the relative differences in dust deposition (as discussed in section 6.4.2).

The observation of comparable (sites 31BB and 17PC) or enhanced (Greenland, site 37BB) dust deposition during MIS 4 relative to MIS 2 is unexpected given traditional models of the climatic parameters driving global dustiness. Of the four hypotheses evaluated to explain this apparent inconsistency, evidence exists in support of enhanced dust supply, suspension and transport (1 and 2). Additional data are needed to constrain the potential role of unidentified dust source areas (3) and changes in the strength of tropical hydrology (4).

6.3 Orbital cyclicity

Previous work has sought to identify connections between the position and/or strength of the ITCZ (Liu et al., 2015) and dust abundance (Larrasoana et al., 2003) with variations in Earth's orbit related to parameters of eccentricity, obliquity and axial precession. In this study, despite more than sufficient sampling to resolve even the highest frequency orbital components, spectral analysis of 37BB and 31BB reveals little power above background at frequencies associated with orbital parameters (Figure 8). The time series from core 17PC does visually demonstrate some power in the 41 kyr and 23 kyr bands, and these periodicities are statistically significant in the power spectral density, but power is not concentrated there. We suspect the absence of strong coherence between orbital periodicities in the cores is because these dust flux records reflect the

combined influence of dust generation, entrainment, transport and possibly periodic ITCZ deposition, and thus either do not respond in a simple way to changes driven by orbital variability, or there are other more dominant controls on their variability.

6.4 ITCZ Movement

6.4.1 Glacial/Interglacial Position

Recent modeling work (eg: McGee et al., 2014) has suggested that long term shifts of the ITCZ are difficult to sustain without significant and persistent perturbations of the cross equatorial heat flux, which are not supported by paleoceanographic reconstructions on glacial-interglacial timescales (Böhm et al., 2014). On these timescales increases and decreases in the magnitude of dust delivered to our three study sites are largely coherent, making it difficult to distinguish the hemispheric origin of dust delivered at each site. However, during MIS 2, from approximately 29 to 14 ka, and during MIS 4 from 71 to 57 ka, site 37BB displays the highest dust fluxes of any of the three sites (Figure 4) despite having the lowest overall sedimentation rates. This observation indicates that the ITCZ likely delivered relatively more precipitation and dust to northernmost site (7.0°N), pointing to a Pacific ITCZ at a latitude similar to its present day position (Figure 1) consistent with work from the Eastern Equatorial Pacific (Xie and Marcantonio, 2012), dust provenance data from the same ML1208 sites studied here (Reimi and Marcantonio, 2016), and modeling results (McGee et al., 2014). Conclusions about the average position of the ITCZ during MIS 3 and MIS 5 are much harder to draw because of the frequency of millennial events in both hemispheres, which are not always well resolved by our records.

6.4.2 Millennial shifts

Previous intercomparison of a time-subset of the three dust flux records presented here identified an abrupt shift of the ITCZ during TII in concert with the North Atlantic HS11

(Jacobel et al., 2016). Here, we examine whether significant features of our new 150 kyr records might also be indicative of ITCZ shifts. Key to our reconstruction is the observation that while global dust fluxes respond in a coherent way to glacial-interglacial variability, millennial dust increases in the northern and southern hemispheres are frequently out of phase, with maximum Antarctic dust fluxes preceding or nearly coincident with the onset of enhanced dust deposition in Greenland (Supplementary Figure 4). An example of the temporal offset between dust fluxes in Antarctica and Greenland is the relationship observed from 95 to 80 ka, when dust fluxes in Antarctica peaked and begin to decline before Greenland dust fluxes increased rapidly in association with GS22 (Figure 6).

6.4.2.1 *Greenland stadials*

Over the last 150 kyr our highest resolution dust flux record, 31BB, records the signatures of abrupt dust flux increases associated with HS11, GS26, GS25, GS24, GS22, most likely GS23 (Jacobel et al., 2016 and our sections 6.2.1 and 6.2.2) and possibly GS18 (discussed below). These Greenland stadials are among the longest duration cooling events and show the largest magnitude of corresponding Antarctic warming and adjustment of the interhemispheric thermal gradient, epitomizing the biopolar seesaw (Figure 9). The modern day position of site 31BB is on the southern edge of the ITCZ and precipitation at 4.5°N is largest during boreal spring (Figure 1) when Asian dust storms are common due to cyclogenesis (Roe, 2009; Serno et al., 2015). Thus, while the observation of Northern Hemisphere dust events in 31BB is not necessarily indicative of a large change in ITCZ position, the presence of NH dust events in 31BB does require an ITCZ shift south of its modern position

During GS24, GS25 and GS26 site 17PC has sedimentation rates very similar to 31BB (2.4 vs 2.7 cm ka⁻¹), however the signature of these events is not easily identifiable at 17PC (Figure

3c). At least two explanations exist for this difference. First, a sedimentary modeling experiment suggests that a mixed layer depth of 11 cm would largely obscure the dust flux signatures of GS24, 25 and 26 at 17PC. This mixed layer depth is twice that simulated for site 31BB and at the upper limit of “diffusive” mixing attributed to benthic activities (Smith et al., 1997). However, we do not have *in situ* data on bioturbation at site 17PC and thus we cannot evaluate this hypothesis. A second possibility is that the contribution of NH dust reaching 0.5°N during these stadials was negligible, indicating that the ITCZ shifted southwards but not far enough to contribute NH dust to 17PC.

An analogous situation may have occurred during GS22 and 23 when multiple peaks of dust are observed at 17PC (Figure 3). During this interval, accumulation rates at 17PC average 2.7 cm ka⁻¹ while at 31BB they are closer to 4.2 cm ka⁻¹. The early part of the increase between 95 and 86 ka probably corresponds with a SH dust flux increase from 95-88 ka (Figure 6e), however we cannot eliminate the possibility that the peak from 86-83 ka is related to GS22/HS8. If the dust flux increase at 17PC is indeed related to GS22/HS8, a substantial shift of the ITCZ southwards would have to have occurred, allowing NH-sourced dust to reach the equator.

6.4.2.2 Marine Isotope Stage 4

In Section 6.2 we discussed a substantial body of evidence indicating dustier and windier conditions during MIS 4. However, the hypotheses explored in that section do not explain the higher dust flux at core 37BB relative to fluxes at sites 31BB and 17PC. The observation of the highest fluxes at the northernmost site is remarkable considering that sedimentation rates at site 37BB (1.2 cm ka⁻¹) are almost three times lower than at 31BB (3.3 cm ka⁻¹) and less than half those at 17PC (2.6 cm ka⁻¹) between 80 and 60 ka (Figure 4c, Supplementary Figure 1). These large differences in accumulation, in combination with bioturbation, would be expected to lead

to smoothing of the signal's amplitude at 37BB relative to the records at the other two sites. Instead, dust fluxes at 37BB are at maximum ~20% higher beginning at ~64.4 ka (allowing for 2 sigma age and analytical uncertainty in our data), a difference that cannot be attributed to more intense bioturbation at the southern sites. We propose that the reason for the substantial difference in dust flux between the northernmost and two more southern sites is that the mean ITCZ was positioned northwards at that time relative to GS26, 25, 24, 23 and 22. An increase in the total amount of dust in the tropical atmosphere alone would not yield the observed pattern because atmospheric dust removal and deposition at all sites would be proportionally larger.

During GS18/HS6 cold stadial conditions characterized Greenland (Figure 4a) while Antarctica experienced warming associated with Antarctic Isotopic Maximum (AIM) 18 (Figure 4e). This asymmetry would have caused a shift in the thermal gradient and may have pulled the ITCZ southwards from a northerly position associated with MIS 5a warmth in the North Atlantic. In sum it appears that the MIS 4 and/or GS18/HS6 position of the ITCZ may represent a southward shift, albeit to a latitude northwards of its position during MIS 5 millennial events. It remains unclear whether that latitude was north, south or similar to the modern day ITCZ.

6.4.2.3 Other millennial shifts

Atlantic ITCZ hydrological shifts in response to North Atlantic HS have been previously observed for HS4, 5 and 6 as recorded by Brazilian speleothems (Wang et al., 2004) and a salinity reconstruction from the eastern equatorial Pacific has been used to suggest decreases in moisture transport in association with HS during MIS 2-4 (Leduc et al., 2007). Given that the duration of these events and magnitude of associated Antarctic warming is in line with other events we do successfully identify (Figure 9), we strongly suspect that dust flux increases associated with HS4 and 5 are represented in our records. Unfortunately, identification of these

events is challenging because the frequency of dust flux increases in the SH and NH (the latter due to GS events) is dramatically increased during MIS 3 relative to MIS 4 and 5 (Figure 4). Given the limitations of our age models, the influence of bioturbation and the sampling resolution of our new records we are thus unable to unequivocally identify these events or to associate them with shifts of the Pacific ITCZ.

Two of the best-studied Heinrich Stadials (HS1 and HS0/YD) also occurred during the interval presented in this paper. Because these events occurred during the last deglacial termination, in the same way that HS11 occurred during the penultimate termination, they are the focus of a separate, detailed comparison study of the last deglaciation (Jacobel et al., in prep). Here, we note that in contrast with the penultimate deglaciation, core 17PC (the southernmost site) shows no evidence of having received NH dust associated with H1 or the YD during TI (Figure 4b and c). However, there is some evidence in 31BB of structure similar to that observed in the Greenland ice core record during both deglacial stadial events. This leads us to conclude that the position of the ITCZ during TI stadial events may have been very similar to its position during GS 24, 25 and 26. This inference stands in contrast with data from TII indicating a more substantial southwards shift of the ITCZ during that deglaciation. Indeed our data suggest that the southward shift of the ITCZ that occurred during TII (Jacobel et al., 2016) has been unparalleled over at least the last 150 kyr.

7. Conclusions

Dust plays a crucial role in the climate system as both a force that can drive climate and as feedback to environmental change. Here we have reconstructed millennial-scale resolution records of dust flux from the tropics, the first to cover MIS5-3. Our latitudinal transect of cores sheds light on both the atmospheric load of dust in the tropics and the relative position of the

ITCZ over the last 150 kyr. These new records highlight the complex role dust generation, deflation, transport and deposition processes play in influencing the tropical record of dust flux. On glacial-interglacial timescales these processes result in a scenario where dusty glacial periods abruptly transition to interglacial periods with ~3 times lower dust fluxes, returning to dustier conditions in step with cooling global temperatures and ice volume growth. Our data suggest that any shifts of the ITCZ in response to glacial-interglacial changes in the interhemispheric thermal gradient were likely to have been small in the central equatorial Pacific, at least over the last interglacial and glacial cycle. On millennial timescales our new records reveal marked changes in dust abundance, and small shifts in ITCZ position relative to modern in response to changes in hemispheric thermal gradients driven by a handful of the longest North Atlantic stadial events. These new data provide evidence that high-latitude cooling and circulation changes have repeatedly influenced the tropical radiation balance via changing dust fluxes and have occasionally caused significant perturbations in tropical hydrology on relatively short timescales. Our inferences strongly support the utility of highly (temporally and spatially) resolved records of paleoclimate, and should encourage the development of high-resolution records from the tropics to constrain other components of the climate system and responses to high-latitude climate forcing.

Acknowledgements

We thank the core repository at LDEO for providing samples and Marty Fleisher for assistance in the lab. Special thanks to Jeremy Hoffman for generously sharing his MonteXCM code to improve the age models presented in this paper and to Jean Lynch-Stieglitz for assistance with and discussion of ^{14}C dates. This research was funded in part by awards from the

Geological Society of America (A.W.J.) and US National Science Foundation AGS-1502889 (J.F.M. and G.W.), and benefited from previous support from NSF OCE-1003374, OCE-1159053 and OCE-1158886, and the Comer Science and Education Foundation.

Data Archiving

Line Islands dust flux data are archived at the National Oceanic and Atmospheric Administration National Centers for Environmental Information (NCEI) database and are also available as a supplement to this manuscript.

Figure Captions

Figure 1. Map of study area. ML1208 core sites 37BB, 31BB and 17PC (black stars) are shown relative to Hawaii (white islands), and mean annual precipitation in mm per day (colors) at left of figure. Inset shows the location of this map (red rectangle) in the broader context of the Pacific Ocean. Panel at right indicates the mean annual distribution of precipitation (black line) at 160°W as well as the long term mean precipitation during the months of January (blue line) and August (orange line). Basemap generated using the global precipitation climatology product (GPCP) version 2.2 and histograms generated using Climate Prediction Center long term mean data, both from National Oceanic and Atmospheric Administration's (NOAA's) Physical Sciences Division, Earth System Research Laboratory. Inset generated in GeoMapApp. Figure and caption modified from Jacobel *et al.*, 2016.

Figure 2. Isotopic abundance and flux data. Mass flux (a), ^{232}Th activity (b) and ^{232}Th flux (c) data from sites 37BB (teal squares), 31BB (blue triangles) and 17PC (purple circles). Lines (matching colors) represent values after the application of a 3-point, 1st order Savitsky-Golay

578 filter.

579 **Figure 3. Dust fluxes.** Dust fluxes and 2 sigma uncertainties for sites 37BB (teal squares), 31BB
 580 (blue triangles) and 17PC (purple circles). Lines (matching colors) represent the average of
 581 replicate points. Radiocarbon control points are indicated at the bottom of the figure in filled
 582 symbols for 37BB and 17PC using the same symbol and color scheme as above. Note that
 583 radiocarbon ages from 31BB were not used to determine the age model for that core. Marine
 584 Isotope Stage boundaries are indicated at the top of the plot with grey shading indicating glacial
 585 stages. Note that all axes have identical scales.

586 **Figure 4. Global dust flux comparison.** Oxygen isotope composition of the Greenland NGRIP
 587 ice core in blue (North Greenland Ice Core Project members, 2004) **(a)**. Greenland dust flux
 588 record from NGRIP in dark green (Ruth et al., 2007) **(b)**. Dust fluxes from the central equatorial
 589 Pacific (this study) with colors as in Figure 3 **(c)**. Dust fluxes from the Antarctic EDC ice core in
 590 light green (Lambert et al., 2012) **(e)**. Deuterium isotopic ratio of the Antarctic EDC ice core in
 591 light blue (Jouzel et al., 2007) **(f)**. Records presented in panels a, b, d and e have been filtered for
 592 clarity (Supplementary Note 6). Marine Isotope Stages labeled as in previous figure with selected
 593 Greenland and Heinrich stadial events labeled following Rasmussen et al., 2014. Arrows in MIS
 594 4 emphasize the relative timing of dust flux decline in each hemisphere. All time series on the
 595 AICC2012 age model of (Veres et al., 2013).

596 **Figure 5. Greenland stadial events 24, 25 and 26.** Dust fluxes from ML1208 31BB in blue **(a)**.
 597 Oxygen isotope data from Greenland (NGRIP) (North Greenland Ice Core Project members,
 598 2004) on the GICC05 age model (Seierstad et al., 2014) (navy blue) **(b)** and AICC2012 (Veres et
 599 al., 2013) (light blue) **(c)** timescales.

Figure 6. Close up of HS8/GS22. Oxygen isotope data from Greenland's NGRIP ice core (dark blue) (North Greenland Ice Core Project members, 2004) **(a)**. Oxygen isotope data from Sanbao Cave, China (Wang et al., 2008) (purple) **(b)**. NGRIP dust flux data (dark green) (Ruth et al., 2007) **(c)**. Dust flux data for core 31BB (blue) **(d)**, EPICA Dome C dust flux data (light green line) (Lambert et al., 2012) and dust flux data from Southern Ocean core TN57-6 (light green dotted line) (Anderson et al., 2014) **(e)**. Deuterium isotope ratio of the Antarctic EDC ice core (light blue) (Jouzel et al., 2007) **(f)**.

Figure 7. Comparison of dust flux records and proxies for wind strength. Dust fluxes for the three ML1208 cores (colors as in previous figures) **(a)**. Iberian margin alkenone-based sea surface temperature record (Pailler and Bard, 2002) (navy) **(b)**. Lake Biwa eolian quartz flux (purple) (Xiao et al., 1999) **(c)**. Marine Isotope Stages labeled as in previous figures.

Figure 8. Power spectral density plots for ML1208 dust flux records. Estimated power spectral density (PSD) for dust fluxes at 37BB **(a)**, 31BB **(b)** and 17PC **(c)** using the Thompson Multitaper Method, normalized to the power of the 100 kyr cycle in each PSD plot. The significance of spectral peaks (solid black lines, dotted black lines 95% confidence intervals) is tested against the null hypothesis of a first order autoregressive process (solid blue line, dotted blue line 95% confidence interval). Solid vertical lines represent frequencies associated with the orbital periodicities of 100, 41 and 23 ka. Peaks of the solid black line above the dotted blue line indicate frequencies with significant concentrations of power at the 95% confidence level.

Figure 9. Stadial events of the last 150 kyr. Cross plot of Greenland stadial duration (ka) and associated Antarctic warming at EPICA Dome C. Events with blue labels have been identified in the ML1208 cores presented in this study. Data and figure adapted from (Marino et al., 2015).

Additional data points added as described in Supplementary Note 7, employing data from (Wang et al., 2008; Masson-Delmotte et al., 2010; Veres et al., 2013; Seierstad et al., 2014).

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