The response of ENSO flavors to mid-Holocene climate: Implications for proxy interpretation

Christina Karamperidou¹, Pedro N. Di Nezio², Axel Timmermann², Fei-Fei Jin¹, and Kim M. Cobb³

¹Department of Atmospheric Sciences, University of Hawai‘i at Mānoa, Honolulu, Hawaii, USA; ²Department of Oceanography, University of Hawai‘i at Mānoa, Honolulu, Hawaii, USA; ³School of Earth and Atmospheric Sciences, Georgia Institute of Technology, Atlanta, Georgia, USA

Abstract The response of El Niño–Southern Oscillation (ENSO) to mid-Holocene boundary conditions remains an open question: paleoclimate proxies and climate model simulations do not agree in the magnitude of the reduction of ENSO variability, while recent proxy evidence from fossil corals from the central Pacific show that the reduction in mid-Holocene ENSO variability compared to the end of the twentieth century is not different from the reduction during other Holocene time intervals. This is inconsistent with the interpretation of lake and ocean sediment records from the eastern Pacific, which show a significant reduction compared to all other Holocene periods. In order to reconcile the seemingly conflicting proxy evidence from the eastern and central Pacific, we hypothesize that ENSO remained active during the mid-Holocene; however, there was a change in the spatial pattern of the sea surface temperature anomalies, also known as ENSO flavors. Using National Center for Atmospheric Research’s Community Climate System Model version 4 forced with mid-Holocene orbital conditions, we find that the frequency of occurrence of the strongest eastern Pacific (EP) events decreases in the mid-Holocene and their variance is reduced by ~30%, while the frequency of central Pacific (CP) events slightly increases and their variance does not change. We also find a shift in the seasonality of EP events, but not in that of CP events.

Lastly, mid-Holocene EP events develop more slowly and decay faster. The differential response of ENSO flavors to mid-Holocene forcing is remotely forced by the West Pacific, where a weakening of the trade winds in early boreal spring in the mid-Holocene initiates an anomalous downwelling annual Kelvin wave, which reaches the eastern Pacific during the ENSO development season, weakens the upper ocean stratification, and results in reduced ENSO upwelling feedback. The simulated reduction in the EP flavor versus the CP flavor in the mid-Holocene is consistent with proxy evidence: The teleconnection patterns of the two flavors with temperature, precipitation, and salinity are distinct, and proxies from different regions of the Pacific might be recording variability associated with only one of the flavors, or some combination of their relative effects.

1. Introduction

Neither models nor observations provide a conclusive answer as to whether El Niño–Southern Oscillation (ENSO) is going to weaken or strengthen in response to greenhouse warming [Meehl et al., 2007; Collins et al., 2010; DiNezio et al., 2012; Cai et al., 2014], or whether ENSO sea surface temperature (SST) anomalies will tend to be located in the central rather than the eastern tropical Pacific [Yeh et al., 2009; Lee and McPhaden, 2010; McPhaden et al., 2011]. Given that ENSO is the dominant mode of tropical variability, the lack of model agreement is an important source of uncertainty for projecting future regional climate change throughout the Pacific basin and beyond [Meehl et al., 2007]. Paleoclimate reconstructions offer the possibility of testing the theories of ENSO response to greenhouse warming, as well as the models used to project this response. The climate of the mid-Holocene—about 6000 years before present—has received much attention owing to proxy interpretations suggesting a significant reduction in climate variability associated with ENSO [Rodbell et al., 1999; Moy et al., 2002; Riedinger et al., 2002; Koutavas et al., 2006; Donders et al., 2008; Conroy et al., 2008].

Climate models of various complexity agree in key features of the response of the tropical Pacific to mid-Holocene orbital forcing [Clement et al., 2000, 2001; Hewitt and Mitchell, 1998; Bush, 2008; Liu et al., 2000; Kitoh and Murakami, 2002; Otto-Bliesner et al., 2003; Brown et al., 2006, 2008]. Most coupled global climate models (GCMs) participating in the Paleoclimate Modelling Intercomparison Projects (PMIP2 and PMIP3) simulate a reduction of ENSO variability, along with a significant reduction in the strength of the annual
cycle of the eastern equatorial Pacific [Masson-Delmotte et al., 2013]. In present-day climate ENSO is strongly influenced by the seasonal cycle of the Pacific cold tongue. Therefore, it is reasonable to expect that the orbitally driven changes in seasonal climate would alter ENSO, as proposed by previous intermediate-complexity or single model studies [e.g., Clement et al., 2000; Sallau et al., 2012].

Seasonal variations in the climate of the Pacific play a key role in the onset and termination of ENSO. First, the strength of the annual cycle of the Pacific cold tongue modulates the strength of the Bjerknes feedback—the positive feedback loop responsible for the growth of ENSO events [Jin et al., 1994, 1996; Tziperman et al., 1994, 1995; Wang and Fang, 1996]. Second, the seasonal migration of the South Pacific Convergence Zone (SPCZ) also plays a role, albeit in the termination of strong El Niño events [Harrison and Vecchi, 1999; Stein et al., 2011; McGregor et al., 2012; Stuecker et al., 2013; Stein et al., 2014]. Changes in processes away from the tropical Pacific could also influence ENSO. For instance, mid-Holocene ENSO could be weakened due to a stronger Asian monsoon through an alteration of the seasonal cycle of the tropical Pacific [Chang et al., 1994; Liu, 2002; Pan et al., 2005; Timmermann et al., 2000, 2000; Brown et al., 2008; Marzin and Braconnot, 2009]. An extratropical mechanism has also been proposed, in which reduced stochastic forcing originating from the North Pacific weakens ENSO [Chiang et al., 2009].

There are, however, key inconsistencies between simulations and paleoclimate reconstructions of mid-Holocene ENSO. First, the magnitude of the reduction in ENSO variability in the models (approximately 10%) is not as large as suggested by some paleoclimate data [Masson-Delmotte et al., 2013; Gagan et al., 2004; Donders et al., 2008]. Terrestrial and marine proxies from the eastern tropical Pacific suggest abrupt changes within the broadly defined mid-Holocene period of 4–7 ka B.P. [Koutavas et al., 2002; Gagan et al., 2004; Donders et al., 2005, 2007; Chazen et al., 2009; Koutavas and Joaides, 2012], while the models simulate gradual changes in ENSO variability [Clement et al., 2000, 2001; Donders et al., 2008; Chiang et al., 2009]. At the same time, the presence of noise, the short length of many of the available proxy records (e.g., corals), and the multiple proxy resolutions (from seasonally to centennially resolved) create considerable uncertainty regarding the magnitude of the suggested change and further complicate model-proxy comparisons [Wittenberg, 2009; Cobb et al., 2013]. At present, it is difficult to reject the hypothesis that internal ENSO variability on decadal and centennial timescales dominates over the forced orbital response during the Holocene [Wolf et al., 2011; Cobb et al., 2013]. In addition, model-proxy disagreement could be due to changes in the ENSO teleconnections due to a northward shift of the climatological Intertropical Convergence Zone (ITCZ) in the mid-Holocene [Woodroe et al., 2003; Gagan et al., 2004; McGregor and Gagan, 2004; Koutavas et al., 2006], and not necessarily due to changes in ENSO variability itself. Last, while eastern Pacific archives indicate that interannual variability in the mid-Holocene at that location is weaker than even the case of completely vanished ENSO [Koutavas and Joaides, 2012], recently available proxy records from fossil corals from the central Pacific show that ENSO’s strength during the mid-Holocene was comparable to that during the last millennium [Cobb et al., 2013].

In this paper, we present a physical mechanism that could help reconcile the seemingly conflicting proxy evidence of mid-Holocene ENSO from the eastern and the central Pacific regions. Our hypothesis is motivated by the studies discussed above, which collectively suggest a weaker ENSO in the eastern Pacific compared to all other Holocene time intervals [e.g., Koutavas and Joaides, 2012] but question the presence of a reduction of similar magnitude in the central Pacific, again as compared to other Holocene time intervals except for the post-1970 era [Cobb et al., 2013]. We hypothesize that the proxy data may reflect a differential response of the spatial pattern of ENSO’s SST anomalies—also known as ENSO flavors—to orbitally driven changes in the seasonal cycle. That is, during the mid-Holocene, ENSO events with SST anomalies concentrated in the eastern Pacific (the Eastern Pacific (EP) flavor) were weaker and/or less frequent, while ENSO events with SST anomalies concentrated in the central Pacific (central Pacific (CP) flavor) were mostly insensitive to orbital forcing.

Various mechanisms have been proposed to explain the existence of ENSO flavors, including the lack of thermocline-surface interactions in the central Pacific—leading to a preponderance of the zonal advection feedback over the thermocline feedback [Yeh et al., 2009]– and of a mechanism of transition to cold events [Kug et al., 2009; Kao and Yu, 2009; Yu and Kim, 2010; Newman et al., 2011a, 2011b], changes in intraseasonal variability and its coupling with low-frequency atmospheric flow [Kug et al., 2009], differences in the location and intensity of westerly winds [Hu et al., 2012], the timing of the onset of SST anomalies [Xu and Chan, 2001], Asian and Australian monsoon forcing [Yu et al., 2009], and subtropical atmospheric forcing [Yu et al., 2010]. On the other hand, the occurrence of EP events has been attributed to the nonlinear evolution of ENSO which
does not require CP and EP El Niño events to be different phenomena [Takahashi et al., 2011], or conversely, to a statistical artifact due to the nonlinearity between El Niño and La Niña [Monahan and Dai, 2004; L’Heureux et al., 2012], and natural variability [Yeh et al., 2011; Newman et al., 2011a]. An investigation of the response of ENSO flavors to orbital forcing could improve our understanding of their origins, especially in light of evidence about their interaction with the annual cycle [McGregor et al., 2013b].

Should our hypothesis find support, it would not only reconcile proxy evidence from around the Pacific, since CP ENSO events have distinct teleconnections from the EP events [Larkin and Harrison, 2005; Hu et al., 2012; Ashok et al., 2007; Kim et al., 2009; Yeh et al., 2009; di Lorenzo et al., 2010; Mo, 2010; Yu and Kim, 2010; Hoerling and Kumar, 2002; Wang and Henderson, 2007; Weng et al., 2007], but would also be consistent with a weaker annual cycle during the mid-Holocene, since CP events do interact with the annual cycle, whereas CP events do not [McGregor et al., 2013b]. In fact, during the preparation of the present paper, our hypothesis gained further support by the study of Carré et al. [2014], who use fossil mollusk shells from Peru and fossil corals from the central Pacific to conclude that a predominance of CP events compared to EP events is possible in the period 6.7 to 7.5 ka B.P.

Here we explore this hypothesis using preindustrial control and mid-Holocene simulations conducted with the Community Climate System Model version 4 (CCSM4), developed by the National Center for Atmospheric Research (NCAR). We provide a new detailed representation of mid-Holocene tropical Pacific climate, which is largely consistent with newly available records of SST variability from paleoclimate proxies. The remainder of the paper is organized as follows: After a brief description of the model simulations (section 2), we describe changes in ENSO flavors in the CCSM4 preindustrial control and mid-Holocene simulations (section 3). In section 4, we present orbitally induced changes in the seasonal cycle of the tropical Pacific, and in section 5, we study the consequent changes in the main feedbacks that control ENSO event development and decay. Section 6 discusses the implications of our results for the interpretation of paleoclimate proxy evidence. Summary and discussion close the paper (section 7).

2. Climate Simulations

In this study, we use NCAR’s Climate System Model version 4 (CCSM4). CCSM4 is a climate model consisting of coupled atmosphere and ocean general circulation models (GCMs) and comprehensive land and cryosphere models. The reader is referred to Gent et al. [2011] for specific information about CCSM4. The preindustrial control simulation (hereafter piControl) analyzed here spans 1300 years and includes interactions between components of the climate system (ocean, atmosphere, cryosphere, and land) configured at nominal 1° latitude-longitude resolution and forced by constant preindustrial (1860) greenhouse gas concentrations. CCSM4 simulates ENSO realistically in the preindustrial experiment, including a 3 to 6 year period, asymmetry between warm and cold events, events with a range of amplitude and return times, and multidecadal
modulation of ENSO [Deser et al., 2012]. As will be shown in section 3, CCSM4 is also capable of simulating realistic SST patterns associated with ENSO’s EP and CP events (Figure 1).

The mid-Holocene simulation branches off year 800 of piControl, spans 500 years, and was performed with the same version of CCSM4, following the experimental protocol of the Paleoclimate Modelling Intercomparison Program 3 (PMIP3). The CO₂ concentration is set to 280 ppm (preindustrial values), the eccentricity is 0.018682 compared to 0.016724 in piControl, the obliquity is 24.105° (piControl is 23.446), and the angle between fall equinox and perihelion is 0.87° compared to 102.04° at piControl, in order to reflect the effect of changes in the Earth’s orbit on insolation at about 6 ka B.P. The resulting changes in seasonal downwelling shortwave radiation are shown in Figure A1 in Appendix A.

3. ENSO Flavors in the Mid-Holocene
3.1. Definition of ENSO Flavors
It has long been recognized that at least 2 degrees of freedom are needed to describe SST anomalies during the evolution of an ENSO cycle [Timmermann, 1999; Trenberth and Stepaniak, 2001]. Some events, like the 1997–1998 event, have SST anomalies localized in the eastern Pacific, the so-called EP ENSO. Other events have maximum SST anomalies located in the central Pacific, called dateline El Niño [Larkin and Harrison, 2005], warm pool El Niño [Kug et al., 2009], central Pacific (CP) El Niño [Kao and Yu, 2009], or El Niño Modoki [Ashok et al., 2007].

Several indices have been used to characterize the two different flavors of ENSO. Ren and Jin [2011] showed that the typical NINO3 and NINO4 indices are not an orthogonal coordinate system to capture the two flavors of ENSO and therefore developed new indices based on combinations of the standard ones. Takahashi et al. [2011] also propose ENSO indices that are effectively a rotation of the first (PC1) and second principal components (PC2) of the tropical Pacific SST anomalies:

\[ E_{\text{index}} = \frac{PC1 - PC2}{\sqrt{2}} \quad C_{\text{index}} = \frac{PC1 + PC2}{\sqrt{2}}. \]  

Figure 1 shows the regression of SST anomalies on the E index and the C index in the CCSM4 piControl simulation and in observations. The indices were computed as per Takahashi et al. [2011]: The principal components of the SST anomalies in the region [110°E–60°W, 10°S–10°N] are based on the full 1300 year climatology, are standardized and passed through a three-point (1-2-1) weighted running average filter. The first two empirical orthogonal function patterns in CCSM4 explain 73.7 and 6.3% of the variance, respectively. The regression patterns in CCSM4 are in good agreement with the observed (explaining 66% and 10%, respectively); note, however, that the simulated SST anomalies are latitudinally more constrained and extend further to the west in the model, which is a known problem in GCM simulations of ENSO anomalies [Bellenger et al., 2014]. Based on the regions of maximum SST variance in Figure 1, the model’s standard ENSO regions are slightly shifted compared to observations [also see Capotondi, 2013]. In this paper, however, we avoid using these standard ENSO indices; rather, we base all our calculations on the E index and the C index (equation (1)).

This decomposition of ENSO flavors presents two distinctive features: (1) the E index captures the strong EP events (see Figures 1a and 1c ) and (2) the C index captures CP El Niño and all La Niña events (Figures 1b and 1d). Figure 2a shows the scatterplot and bivariate probability density function of the two leading principal components of monthly SST anomalies (PC1 and PC2) for piControl, averaged over October–April. The model simulates the nonlinearity in the relationship between PC1 and PC2 identified by Takahashi et al. [2011] in observations. Superposed are October–April averages for events with the E index larger than 2 standard deviations from zero (solid black circles), and October–April averages for events with C index larger than 1 standard deviation away from zero (gray-filled circles). Strong El Niño events lie along the E index axis, while moderate warm El Niño and La Niña events lie along the C index axis.

3.2. Response to Mid-Holocene Orbital Forcing
Strong EP events (solid circles) are significantly reduced in number in the mid-Holocene simulation (Figure 2b), although they do not completely disappear. The October–April variance of the E index reduces from 0.72 in piControl to 0.52 in mid-Holocene (a 30% decrease). In contrast, the variance of the C index does not change between the two climates and is equal to 1.03. The reduction in EP events is more clearly seen in the difference of the bivariate probability density functions (pdfs) between the two climates (Figure 2c). The reduction in the pdf mass at the tails along the E index axis (blue dashed contours) indicates that the frequency of occurrence
of the largest EP events decreases. Indeed, the average frequency of EP events decreases from 3 per century in the piControl simulation to 1.8 per century in the mid-Holocene simulation. On the contrary, the pdf along the C index does not change appreciably (Figure 2c), indicating that these types of events do not respond to mid-Holocene forcing. The average frequency of CP events is 10 per century in piControl and 12 per century in mid-Holocene.

Figure 3 shows the time-longitude plots of SST anomalies for the composite EP and CP events in piControl and mid-Holocene, as well as their difference (Figures 3c and 3f). Anomalies are computed with respect to each simulation’s climatology. The composites include 38 piControl EP events, 9 mid-Holocene EP events, 130 piControl CP events, and 63 mid-Holocene CP events. The events used in the composites are shown in black (EP) and gray (CP) circles in Figure 2. Statistical significance for the difference in SST (Figures 3c and 3f) is assessed as follows: Of the total of 47 EP events in both climates, we randomly select 38 and 9 (without replacement) to be termed piControl and mid-Holocene EP events, respectively. The difference time-longitude matrix is computed. We repeat this process 1000 times and compute the statistics of the resulting difference matrices. We define the statistical significance level at the 5th and 95th percentile. The same process is followed for the CP events. For the EP events, the significant differences lie within $[-0.42, 0.42]$ (stippled regions in Figure 3), and for the CP events they lie within $[-0.18, 0.18]$.

In terms of the magnitude of events, EP events have almost the same magnitude—as measured by the standard deviation of the NINO3 index—at approximately 3°C in both climates (also seen in Figures 3a and 3b). The main difference, in addition to the reduction in the frequency of EP events in the mid-Holocene, is in the length and timing of the development and decay phase. The development of EP events in the mid-Holocene is slow and starts more than 18 months before the peak of the event. The peak is delayed by approximately 2 months into February–March. The decay of EP events in the mid-Holocene is steeper, and the system moves into a cold phase within 4 months after the peak. CP events (Figures 3d–3f) also start developing earlier in the mid-Holocene and are slightly more intense (by 0.6–0.8°C). There is a westward propagation of SST anomalies in the CP events, which is also enhanced in the mid-Holocene. We will show below that the shift in the development and decay of EP events in the mid-Holocene can be attributed to the a change in the seasonality of the peak of the events which is a consequence of changes in the annual cycle of the tropical Pacific.

Figure 4a shows the percentage of winter (October–April (ONDJFM)) months with EP event peaks by month and climate. The identification of peak months is done as follows: We select the 36 month period centered at December–February of each year in EP status as highlighted by black points in Figures 2a and 2b. The peak month is defined as the month within this period in which the $E$ index peaks. In piControl, the majority (40%) of EP events peak in December, with April and February following with 29% and 24%, respectively. In mid-Holocene, this percentage drops dramatically, with only 7% of events peaking in December, while
the majority peak in February (46%) and April (38%). These results indicate a shift in the seasonality of EP events toward late winter and early spring (February–April) in the mid-Holocene. Figure 4b shows the same calculation for CP events. The most notable change is the drop in percentage of CP peaks in April (from 21% in piControl to 7% in mid-Holocene). Increases in the percentage are found for October, November, December, and March, ranging from 2% to 8%. However, the distribution of peak months is not significantly changing, as in the EP case.

Figure 3. Composite SST anomaly plots for (a–c) EP and (d–f) CP events in piControl (Figures 3a and 3d), mid-Holocene (Figures 3b and 3e), and their difference (Figures 3c and 3f). Stippled areas in Figures 3c and 3f denote statistical significance. All fields are averaged between 5°S and 5°N. Anomalies are computed with respect to each simulation's climatology.

Figure 4. The percentage of winter (ONDJFMA) months with (a) EP event peaks and (b) CP event peaks by month and climate.
Figure 5. Climatological mean monthly SST (shading), precipitation (contours), and winds (vectors) during boreal summer/fall (JASO) and boreal winter/spring (FMAM) and in (a, b) piControl, as well as the change in the (c, d) mid-Holocene.

In summary, CCSM4 simulates differential responses of the EP and CP flavors of ENSO to orbital forcing. The three main features of this response are as follows:

1. The frequency of occurrence of EP events decreases, whereas that of CP events slightly increases.
2. The peak of EP events shifts by about 2 months, from December in piControl to February in the mid-Holocene. The peak of CP events does not shift.
3. The EP events decay faster in the mid-Holocene climate compared with piControl by approximately 2 months. CP events terminate faster in the mid-Holocene.

In the following sections we will show that these main features are a consequence of changes in the seasonality of tropical Pacific SSTs and winds. These changes in the seasonal cycle in turn influence the onset and termination of EP and CP events differentially, resulting in weaker EP ENSO during the mid-Holocene.

4. Seasonal Changes in Tropical Pacific Climate

In preindustrial climate, CCSM4 simulates a fully developed cold tongue during late summer/fall (July–October (JASO)), along with stronger SE and NE trade winds, and a maximum northward extent of the ITCZ (Figure 5a colors, vectors, and contours, respectively). Consistent with observations, the simulated cold tongue vanishes during late boreal winter/spring (February–May (FMAM)) along with a slowing down of the SE trades and a weaker ITCZ (Figure 5b).

The changes in the orbital parameters during the mid-Holocene result in increased insolation in the tropics in JASO (with a peak amplitude change in September compared to piControl) and decreased insolation during FMAM, as shown in Figure A1 in Appendix A. The tropical climate response to these changes in insolation is characterized by a weakening of the seasonal cycle in the cold-tongue region, with warmer SSTs there during JASO and colder during FMAM (Figures 5c and 5d, colors). In contrast, both the NE and SE trades strengthen during JASO (Figure 5c, vectors)—the season when they are seasonally stronger—and they weaken in FMAM (Figure 5d, vectors)—the season when they are seasonally weaker. The response of the SE and NE trade winds does not follow the changes in the cold tongue. Their response could be related to changes in the subtropical highs, which strengthen forced by local and remote changes in diabatic heating associated with the monsoons [Mantsis et al., 2013a]. The stronger trade winds during JASO explain the widespread cooling of the tropical Pacific due to increased evaporative cooling (excluding the cold tongue that warms), offsetting the warming due to increased insolation (see Figure A1c in Appendix A). The tropical Pacific also exhibits widespread cooling during FMAM, however, in this case due to decreased insolation (Figure A1d in Appendix A).
Ocean dynamics must be invoked to further explain the simulated response of the cold tongue, which warms during JASO despite stronger trade winds and cools during FMAM despite of the weaker trade winds. Figure 6a shows the time-longitude plot of the difference in climatological trade wind stress in the tropical Pacific between the two climates. Anomalous westerly winds in the west Pacific in FMAM result in an anomalous “annual Kelvin wave,” which is seen in the difference in thermocline depth between mid-Holocene and piControl (Figure 6b). This downwelling Kelvin wave (also seen in the difference of surface ocean current velocity $u_{os}$ in Figure 6c) is initiated during boreal spring in the West Pacific, propagates eastward, reaches the Eastern Pacific during boreal fall, and is responsible for a reduction of the stratification $dT/dz$ in the eastern Pacific.

The seasonal evolution of the thermal stratification, measured by the difference between SST and ocean temperature at 50 m depth ($T_s - T_{sub}$), shows anomalous eastward propagation in the mid-Holocene associated with the annual Kelvin wave (Figure 6d). Over the eastern Pacific, CCSM4 simulates increased $T_s - T_{sub}$ in boreal winter and early spring during, with a maximum in January–February. This stratification increase in the eastern Pacific results in enhanced vertical advection of colder subsurface waters, which acts to cool
the seasonally warmer SSTs (see FMAM in Figure 5). Conversely, $T_s - T_{sub}$ decreases during boreal fall (6d), resulting in decreased vertical advection of colder waters, which acts to warm the seasonally colder SSTs (see JASO in Figure 5). These remotely forced stratification changes manifest as a weakening of the annual cycle of cold-tongue SSTs in the mid-Holocene.

5. Seasonal Controls of ENSO Flavor Response to Mid-Holocene Forcing

5.1. The Role of Stratification in the East Pacific

The main differences in ENSO flavors between the two climates include a slower development and faster decay of EP events in the mid-Holocene, as well as a shift in their peak by approximately 2 months (section 3 and Figure 3). In this section, we show that these differences result from changes in the main ENSO feedback mechanisms driven by the orbitally driven changes in the seasonal cycle discussed in section 4.

We performed a heat budget analysis of CCSM4 output to diagnose the physical mechanisms involved in the ENSO changes described above. The heat budget consists of the anomalous heat content tendency $Q'_t = \rho_0c_p \int_0^H \frac{\partial T}{\partial t} dz$, which is related to the tendency of (i.e., growth and decay) of interannual SST anomalies $T'$. ($\rho_0$ is a reference density of sea water, $c_p$ is the specific heat of sea water, $H$ is the thickness of the constant-depth layer over which the terms are integrated). $Q'_t$ is approximately balanced by the anomalous advection of temperature by ocean currents ($Q'_{adv}$) and by the net air-sea heat flux ($Q'_{atm}$). The dominant contributions to $Q'_{adv}$ during development and decay phases of El Niño events are the zonal advection feedback $Q'_{za} = -\bar{\rho}c_p \int_0^H \frac{\partial T}{\partial x} u' dz$, the thermocline feedback $Q'_{tc} = -\bar{\rho}c_p \int_0^H \frac{\partial T}{\partial z} w' dz$, and the upwelling feedback $Q'_{uw} = -\bar{\rho}c_p \int_0^H \frac{\partial T}{\partial z} \frac{\partial T}{\partial z} dz$, where $\bar{T}$ is the climatological monthly mean temperature. These feedbacks can be quantified via an ENSO heat budget analysis, which is computed following the methodology of DiNezio and Deser [2014]. Details may be found in Appendix A.
Figures 7a–7c show the time-longitude composite of total heat content tendency \( Q'_t \) for EP events in (a) piControl, and (b) mid-Holocene, as well as (c) their difference. As expected, the development and decay of SST anomalies shown in Figure 3 closely follows the total heat content tendency. In both climates the upwelling feedback \( Q'_{uw} \) explains most of the \( Q'_t \) of EP events over the NINO1+2 and NINO3 regions—the center of action of the EP events (Figures 7d–7f). \( Q'_{tc} \) explains the remaining \( Q'_t \) (approximately 40–60 W m\(^{-2}\)) (not shown). The \( Q'_t \) changeduring themid-Holocene is also entirely explained by the change in \( Q'_{uw} \) (compare Figure 7c and Figure 7f). \( Q'_{uw} \) and \( Q'_{tc} \) exhibit a smaller magnitude in the mid-Holocene consistent with the slower development of the events (Figure 3). Moreover, \( Q'_{uw} \) and \( Q'_{tc} \) become negative more sharply in mid-Holocene than in piControl, consistent with a faster termination of EP events.

What causes the weakening of the upwelling feedback in the mid-Holocene? The change in upwelling feedback (\( \Delta Q'_{uw} \)) is predominantly due to the change in the background stratification \( \frac{\partial \Delta T}{\partial z} \). In other words it is approximated by

\[
\Delta Q'_{uw} \approx -\rho_0 c_p \int_{-H}^{0} \nu' \frac{\partial \Delta T}{\partial z} \, dz
\]

(2)

where \( \Delta \) indicates the difference between mid-Holocene and piControl. Note that we ignore other contributions to the upwelling feedback that may arise from the convolution of changes in ENSO and in background climatology. We refer the reader to Appendix A for a detailed justification of this approximation.

In present-day climate ENSO events develop from April to October (i.e., their tendency is largest during these months). This is the time of the year when CCSM4 simulates weaker climatological stratification (\( \frac{\partial \Delta T}{\partial z} \)) during the mid-Holocene, represented by \( T_s - T_{sub} \) in Figure 6d. Therefore, the upwelling feedback, which is dominant in the eastern Pacific, becomes weaker during the development phase of ENSO (Figure 6d). Reduced upwelling feedback is also evident in the same the development phase in the EP composite (Figure 7f). We therefore

Figure 8. Composites of CP event (a–c) total heat tendency and (d–f) tendency due to the zonal advection feedback in piControl (Figures 8a and 8d), mid-Holocene (Figures 8b and 8e), and their difference (Figures 8c and 8f). All fields are averaged between 5\(^\circ\)S and 5\(^\circ\)N. Stippled areas in Figures 8c and 8f denote statistical significance.
argue that this weakening of the upwelling feedback is the main mechanism whereby EP events become less frequent or weaker during the mid-Holocene. Conversely, CP events are insensitive to the seasonal shifts in stratification because they are primarily governed by the zonal advection feedback (Figure 8). The peak difference in zonal advection feedback in the central Pacific occurs in May—June of the year preceding the peak (Figure 8f), which coincides with the “annual downwelling Kelvin wave” reaching those longitudes (Figure 6c).

5.2. The Role of Climatological Winds in the West Pacific

CCSM4 simulates EP events that decay faster in the mid-Holocene (Figures 3c and 7c). One of the mechanisms proposed to explain the termination of strong EP events involves the shift of ENSO’s westerly wind anomalies off the equator into the Southern Hemisphere following the seasonal migration of the SPCZ [Harrison and Vecchi, 1999, 2003, 2006; Lengaigne et al., 2006; McGregor et al., 2012]. With this southward shift, the westerly wind anomalies are no longer able to force the oceanic equatorial waveguide, which eventually leads to the termination of strong EP events.

CCSM4 simulates this mechanism realistically in piControl as seen in the composite of wind stress (vectors) and wind stress curl (contours) anomalies (Figure 9a). This shift also occurs in the mid-Holocene simulation, albeit it happens earlier, as seen by comparing Figures 9a and 9b. The earlier shift is clearly seen in the difference composite as a dipole feature (Figure 9c, indicated by the thick black line connecting the stippled—i.e., statistically significant—areas). This stronger and earlier southward shift of wind stress curl anomalies in mid-Holocene would result in an earlier and stronger termination of the EP events, as seen in Figure 3c and 7c.

What is the cause of this earlier shift? We associate this feature again with the shift in peak EP month. EP events primarily peak in January–April in mid-Holocene. In these months, the climatological wind stress in the Southern Hemisphere is stronger (Figure 10b; eastward difference indicated weaker climatological winds). This is associated with a stronger South Pacific Convergence Zone (SPCZ), which can be seen in the difference
Paleoclimatic records in the eastern Pacific show a reduction of mid-Holocene ENSO variance compared to the late twentieth century (see Donders et al. [2008], for a comprehensive review). However, newly available coral records from the central Pacific exhibit variability that is not significantly different from other time intervals of the Holocene that lack significant anomalous orbital forcing [Cobb et al., 2013]. These findings shed some doubt on the hypothesis that ENSO responds to orbital forcing and leave open the possibility of a highly naturally variable ENSO throughout the Holocene. In the previous sections, we provided modeling support for an alternate hypothesis, namely, that ENSO indeed responds to orbital forcing yet this response is different for its two flavors. Thus, proxies from the eastern and central Pacific might be reflecting these distinct responses of the ENSO flavors, as they are simulated by CCSM4.

In modern climate EP El Niño events show SST anomalies in the eastern side of the basin (Figure 11a), and associated precipitation anomalies also shifted toward the east (Figure 11c). Note that only these events drive increased rainfall over the western coast of South America. If these events weaken and become less frequent—as shown by CCSM4 for the mid-Holocene—then it is reasonable to expect that hydrological proxies from the eastern Pacific and equatorial South America will capture reduced “ENSO” variability. This is consistent with proxy evidence from lake sediment records from southern Ecuador and the Galapagos Islands [Rodbell et al., 1999; Moy et al., 2002; Riedinger et al., 2002]. Lake sediment records capture fewer El Niño-related flood events in the mid-Holocene in Ecuador [Rodbell et al., 1999; Moy et al., 2002] and Galapagos [Riedinger et al., 2002; Conroy et al., 2008]. Furthermore, a marine sediment record off Galapagos Island (1.13°S, 89.4°W) also exhibits a drop in foraminifera population variance (reflective of a decrease in annual and/or interannual SST extremes) that has been interpreted as weaker ENSO [Koutavas et al., 2006]. Sandweiss et al. [1996, 2001] also interpret the presence of certain mollusk species in geoarcheological
Figure 11. Observed (a, b) sea surface temperature (SST), (c, d) rainfall, and (e, f) sea surface salinity (SSS) anomalies associated with the central (CP) and eastern Pacific (EP) flavors of ENSO. These patterns are the regression of the SST, rainfall, and SSS anomalies on the $E$ and $C$ indices, respectively. Stippling indicates regressions that are not statistically significant ($p < 0.33$). Note that the precipitation color scale is not linear. Observations covering the 1979–2009 period were used. SST data are from Hadley Centre Global Sea Ice and Sea Surface Temperature [Rayner et al., 2003], rainfall data are from GCPCv2 [Adler et al., 2003], and SSS is Delcroix’s gridded observational data set [Delcroix et al., 2011]. Circles indicate the location of proxies of mid-Holocene ENSO variability.

Evidence along the Peruvian coast as an absence of ENSO variability prior to 5.8 ka B.P. The interpretation of the records from the eastern Pacific is complex due to the nonlinear nature of the runoff proxies (J. Emile-Geay and M. Tingley, Inferring climate variability from nonlinear proxies. Application to paleo-ENSO studies, submitted to Paleoceanography, 2014); sensitivity of the archive to seasonal-interannual variability [e.g., Koutavas and Joanides, 2012]; highland (garúa) versus lowland (convective) precipitation signals recorded by Galapagos lake records [e.g., Trueman and d’Ozouville, 2010; Wolff, 2010]; and spatially heterogeneous nature of the climate signal [e.g., Liu et al., 2013]; however, all proxy types point to a reduction in ENSO-related climate variability. Furthermore, all these records are from locations where EP events have a hydrological (Figure 11c) and SST (Figure 11a) signature uniquely different from CP events; therefore, they could be explained by a reduction in frequency/amplitude of EP events during the mid-Holocene as seen in the CCSM4 simulation.
On the other hand, several δ¹⁸O records spanning the mid-Holocene have been recently obtained from corals from the Christmas (2°N, 157°W) and Fanning (4°N, 160°W) Islands [Cobb et al., 2013]. These records exhibit a reduction in ENSO variability in the mid-Holocene (4–6 ka B.P.), as measured by the standard deviation of the 2–7 year band pass-filtered δ¹⁸O record, of 60% in Christmas Island and 40% in Fanning Island compared to modern coral δ¹⁸O time series spanning the 1968–1998 period. While the variability is less than the observed during the aforementioned recent period, the level of variability during the mid-Holocene is comparable to that of many multcentury periods, including the last millennium. These proxies are thought to be able to record central Pacific El Niño events, as determined by calibration of modern coral δ¹⁸O to the NINO3.4 index [Cobb et al., 2003; Nurhati et al., 2011]. In this region, El Niño events decrease coral δ¹⁸O due to the combined temperature effect of warm SST and δ¹⁸O-depleted rainfall. Our analysis of the observations shows that the EP and CP events have distinct signature on SST, precipitation, and SSS in this region: CP events have a dominant impact on SST (Figure 11b) and precipitation (Figure 11d), while EP events have a dominant impact on SSS (Figure 11e). This contrast is more marked in Christmas, which is closer to the equator and could explain why coral δ¹⁸O records from these islands exhibit different sensitivities. Thus, a weaker EP ENSO during the mid-Holocene could lead to a reduction in δ¹⁸O variability, as shown by these records, but primarily via the salinity effect that this type of events have on this region. This reduction in δ¹⁸O variability could occur even if the CP events did not change.

Other δ¹⁸O records from the central Pacific show greater reduction in ENSO-related variability compared to the twentieth century. McGregor et al. [2013a] present a 175 yearlong, monthly resolved oxygen isotope record, obtained from Christmas Island and dated at around 4.3 ka B.P., which shows that ENSO variance was persistently reduced by 79%, compared to present day. In this study, the comparison with present climate is based on a stacked modern coral record from the same region, spanning 1939–2007. This great reduction in interannual variability is consistent with the 60% reduction reported in Cobb et al. [2013] and can be similarly explained; the difference of comparison periods notwithstanding. As discussed above, due to its location, this proxy would have mixed SST and SSS signals from both EP and CP events (Figure 11): one would expect a mainly CP signature and a less important EP signature in SST, as well as an EP signature in SSS.

McGregor et al. [2013a] also found that circa 4.3 ka B.P. El Niño events peaked 2 months later, in agreement with our analysis of CCSM4, assuming that mid-Holocene conditions relatively hold in this later period. Based on CCSM4’s simulation of mid-Holocene climate, we showed that this shift could be caused by seasonal changes in stratification in the eastern Pacific, which are driven by changes in the trade winds over the western basin and communicated to the eastern basin by a downwelling Kelvin wave. In contrast, McGregor et al. [2013a] argue that the shift is driven by a strengthening of the cross-equatorial winds during the boreal summer and early fall, enhancing eastern equatorial Pacific upwelling, which in turn enhances the zonal equatorial SST gradient, enhances the trade winds, and suppresses El Niño development; this mechanism is not compatible with our modeling analyses. McGregor et al. [2013a] also report an enhancement of the variability in their coral record within the annual band; however, the model simulation does not support a big change in the annual cycle in the central Pacific region (see Figures 5c and 5d). Rather, the weakening of the annual cycle in the eastern Pacific/cold-tongue region is a robust response in the PMIP models [Masson-Delmotte et al., 2013].

In the warm pool region, fossil coral records in Papua New Guinea (PNG) show reduced variability on ENSO timescales. One of the records from Muschu Island (3.25°S, 143°E) spanning the period 7.6–5.4 ka B.P. exhibits an approximate 40% and 15% reduction in ENSO frequency and amplitude compared to late twentieth century, respectively [McGregor and Gagan, 2004], while the record from Huon Peninsula (6.5°S, 147.5°E) also exhibits similar reduction in ENSO variability in the period around 6.5 ka B.P. [Tudhope et al., 2001]. A reduction of 20% in SST variability and 70% in precipitation variability at ENSO periods is reported by Gagan et al. [2004], who use records from the Great Barrier Reef and PNG. The western Pacific is a region where El Niño events have the opposite signature on coral δ¹⁸O than in the central Pacific, owing to cold SST anomalies and reduced rainfall during El Niño events. However, the observed SST patterns for both flavors are very small there (Figures 11a and 11b). In contrast, the rainfall/salinity anomalies could dominate the δ¹⁸O signals in some sites. Moreover, the EP and CP events exhibit distinct rainfall/SSS signatures there. EP events are associated with dry/saltier conditions off the coast of PNG (Figures 11c and 11e). CP events, in contrast, have a rather muted hydrological response because the nodal line of the rainfall anomalies straddles the coast of PNG (Figures 11d and 11f). Note that the SSS anomalies are shifted to the west of the rainfall anomalies for both types of events. This is because the South Equatorial Current advects the associated freshwater flux westward, shifting the SSS anomalies. As a result, the PNG proxies are likely to capture the saltier conditions associated with EP
events. The CP events are characterized by rainfall anomalies shifted farther west than for EP events. The PNG sites fall just in the nodal lines of the SSS pattern; therefore, the PNG corals could also be insensitive to the hydrological signature of CP events. In summary, the total signals recorded in the warm pool sites are likely a mix of salinity and SST signals with influences with varying strength from the two flavors; it is therefore possible that these sites cannot provide further insight into the response of ENSO flavors to mid-Holocene forcing.

To conclude, the reduction in ENSO recorded by the proxies in the eastern Pacific and South America could be due to the significant reduction in ENSO’s EP flavor, shown in section 3. Conversely, the CP flavor remained active and only slightly enhanced in terms of its strength and frequency according to the model, which is consistent with the Fanning and Christmas Island records of Cobb et al. (2013). The central Pacific ENSO proxies could still be recording partially an EP signal, via the salinity effect, as discussed above. Therefore, the simulated shift in the peak month of EP events a couple of months later in the mid-Holocene, which can be explained by the climatological changes in SST, precipitation, and wind stress resulting from orbital forcing, is in agreement with proxies from Christmas Island by McGregor et al. (2013a).

It should be noted that focus of the present paper is to provide modeling support for the idea of differential response of ENSO flavors to orbital forcing, and qualitatively compare the modeling results to available proxy records. In the studies reported above, the definition of the mid-Holocene period as well as the reference period used to infer mid-Holocene ENSO changes varies. However, the primary question that is addressed here is whether ENSO responds to orbital forcing, i.e., whether there are significant changes between the mid-Holocene period and other periods that lack significant anomalous orbital forcing, such as the preindustrial one. Proxy records from the eastern Pacific [e.g., Koutavas and Joanides, 2012] answer this question positively, while records from the central Pacific [e.g., Cobb et al., 2013] answer this question negatively or without ruling out the null hypothesis of ENSO insensitivity to orbital forcing. It is therefore appropriate in the present study to compare the mid-Holocene to the preindustrial model simulation, i.e., a simulation with anomalous orbital forcing to a simulation with present-day orbital forcing, and attempt to discuss the proxies in this context rather than compare to historical or late twentieth century simulations which are forced externally by greenhouse gases among other forcings.

7. Conclusions
Motivated by seemingly conflicting evidence from paleoclimate proxies from the eastern and central Pacific, we studied the response of the two ENSO flavors, eastern and central Pacific El Niño, to orbital forcing, using long simulations of preindustrial and mid-Holocene climate from NCAR’s CCSM4 model. We found a differential response of the two flavors, which we attributed to changes in the seasonality of the cold tongue, and the resulting changes in the heat budget during El Niño events of each flavor in the two climates. Our main findings for the EP flavor can be summarized as follows:

1. The frequency of occurrence of EP events significantly decreases in the mid-Holocene (by 50%). The variance of the EP event index decreases by 30%.
2. The development of EP events in the mid-Holocene is slower, and their decay is faster compared to the preindustrial climate.
3. There is a shift in the seasonality of the EP events, as their peak is delayed by approximately 2 months in the mid-Holocene.
4. The determining factor for the development of strong EP events is the upwelling feedback in the eastern Pacific which is modulated by seasonal changes in stratification. In the mid-Holocene, remote wind forcing in the western Pacific deepens the thermocline over the NINO1+2 region during the boreal fall. The reduced stratification weakens the upwelling feedback resulting in weaker and less frequent EP events.
5. The faster decay of EP events in the mid-Holocene is associated with a stronger and earlier southward shift of wind stress curl anomalies in mid-Holocene, which is in turn associated with weaker climatological wind stress in boreal winter/early spring in the Southern Hemisphere and a stronger SPCZ.

For the CP flavor, we found that

1. The frequency of occurrence of CP events slightly increases in the mid-Holocene (from 10 to 12 per century on average), while the variance of the CP event index remains unchanged.
2. CP events are stronger in the mid-Holocene.
3. There is no shift in the seasonality of CP events.
4. The dominant feedback term for the development of the CP events is the zonal advection feedback, which is stronger in the mid-Holocene possibly in connection to the downwelling Kelvin wave which forms in early spring and reaches the eastern Pacific in late summer to early fall.
5. The mechanism for termination of EP events which involves the southward shift of wind stress anomalies with the progression of seasons does not play a significant role for CP events.

In summary, the differences in the development and decay of both ENSO flavors between the preindustrial and mid-Holocene climate is related to changes in the seasonality of the trade winds. These changes are initiated over the western Pacific and communicated by the ocean to the eastern Pacific. The proposed mechanism to explain the response of ENSO to orbital forcing is presented schematically in Figure 12.

The simulated reduction in the EP flavor and not the CP flavor in the mid-Holocene is consistent with evidence from paleoclimate proxies from the eastern and the central Pacific. The teleconnection patterns of the two flavors with temperature, precipitation, and salinity are distinct, and proxies from different regions in the Pacific might be recording variability of only one of the two flavors, or various combinations of their relative effects. Our model-based analysis suggests that the great reduction in ENSO variability inferred by proxies in the eastern Pacific may be due to a reduction in the EP flavor. On the contrary, the absence of significant reduction in variability in the central Pacific compared to periods lacking anomalous orbital forcing is consistent with the model results that show no significant changes in the CP flavor. Some reduction in the variability inferred by central Pacific ENSO proxies is still consistent with the model results since it could be due to a mixed signal in temperature and salinity from both flavors, with the CP flavor dominating in the temperature effect and the EP flavor dominating in the salinity effect. The issue of mixed signals is particularly burdening in the case of the available western Pacific proxies, therefore, one should be cautious in interpreting them in connection to the two ENSO flavors. It should also be noted that the paleoclimate proxies, and especially the short coral segments, could be subsampling periods of naturally occurring variability in ENSO flavors, which could be superposed on their response to orbital forcing (enhancing or attenuating the latter). For example, one cannot rule out the possibility that the records of Cobb et al. [2013] are sampling a mid-Holocene period of enhanced EP and CP activity superposed on an otherwise significantly muted ENSO activity background, which could result in a signal reduction that is not different from other Holocene periods. Similarly, the records of McGregor et al. [2013a] could be subsampling from a period of naturally decreased EP and CP activity on
top of an orbitally induced ENSO activity reduction, which could result in the greater reduction they report in their study.

This study is possible thanks to the improved realism in the simulation of ENSO in state-of-the-art climate models. The length of the CCSM4 simulations (1300 years of piControl and 500 years of mid-Holocene) allows us to compute robust statistics and make meaningful interclimate comparisons of ENSO variables, especially at the tails of their distributions. CCSM4 is considered one of the best climate models in terms of simulation of ENSO [Deser et al., 2012], as well as its flavors (Figure 1). However, CCSM4 exhibits biases common to other models, such as the well-known “double-ITCZ” and “cold-tongue” biases, which could influence our results. The cold-tongue bias is characterized by a westward extension and lower temperatures of the cold tongue compared to observations [Bellenger et al., 2014], while the double-ITCZ bias is characterized by excessive precipitation over much of the Tropics (e.g., Northern Hemisphere ITCZ, South Pacific Convergence Zone, Maritime Continent, and equatorial Indian Ocean), and insufficient precipitation over the equatorial Pacific, which may lead to overly strong trade winds, excessive latent heat flux, insufficient shortwave radiation flux, and cold SST biases [Lin, 2007]. The cold-tongue bias could favor the occurrence of CP events at the expense of EP events, while the double-ITCZ leads to an unrealistic semiannual cycle of the cold tongue, which could influence the seasonal controls on ENSO flavors. However, there are two other biases common to the best climate models that could make the mechanism proposed here more prominent in the real world. First, CCSM4 simulates an annual cycle that is weaker than observed; therefore, a much larger weakening of the annual cycle would be expected for the mid-Holocene. Second, CCSM4 overestimates the amplitude of ENSO SST variability by approximately 30%; thus, if CCSM4’s ENSO was weaker, then the simulated mid-Holocene reduction in EP ENSO (30%) could potentially lead to a complete disappearance of EP ENSO in the mid-Holocene. After decades of research, the mid-Holocene is still a challenging target for the simulation of ENSO in climate models. Future progress understanding the mid-Holocene ENSO requires improvement of these biases, along with new proxy records from undersampled regions, which include the west, central and eastern equatorial Pacific.

Appendix A: Interpreting Changes in the ENSO Heat Budget

The methodology used here to estimate the heat budget terms has been used to study ENSO dynamics in coupled GCMs [DiNezio et al., 2009, 2012; Capotondi, 2013; DiNezio and Deser, 2014]. The most important feature of this methodology is that a nearly balanced heat budget can be obtained using monthly mean three-dimensional velocity \((u, v, w)\) and temperature \((T)\) fields. Our analysis of the full heat budget shows that for interannual anomalies, the heat budget can be approximated by

\[
Q'_i = \rho_0 c_p \int_{-H}^{0} \frac{\partial T}{\partial t} \, dz = -\rho_0 c_p \int_{-H}^{0} \left( u \frac{\partial T}{\partial x} + v \frac{\partial T}{\partial y} + w \frac{\partial T}{\partial z} \right) \, dz + Q'_{\text{atm}} \tag{A1}
\]

The definitions of the variables in (A1) follow the convention, where primed variables are anomalies with respect to the climatological monthly mean seasonal cycle. Here anomalies are computed with respect to each simulation’s (mid-Holocene and piControl) climatology. The right-hand side of (A1) is the heat storage rate. The first term in the left-hand side is the advection of temperature by ocean currents. In our analysis in section 5, we neglect several temperature advection terms, such as meridional advection and most nonlinear terms, because their interannual variability is small. We focus on the three main feedbacks in the NINO3.4 and NINO1+2 regions, i.e., the zonal advection feedback \(Q'_{za} = -\rho_0 c_p \int_{-H}^{0} \left( u \frac{\partial T}{\partial x} \right) \, dz\), the thermocline feedback \(Q'_{tc} = -\rho_0 c_p \int_{-H}^{0} \left( w \frac{\partial T}{\partial z} \right) \, dz\), and the upwelling feedback \(Q'_{uw} = -\rho_0 c_p \int_{-H}^{0} \left( w'^2 \frac{\partial T}{\partial z} \right) \, dz\). Both the tendency and advection terms are integrated over a constant-depth layer of thickness \(H = 90\) m, which is taken to be 20 m below the base of the ocean mixed layer (similar results are obtained using 10 to 30 m). \(T'\) is the ocean temperature anomaly averaged over depth \(H\), and \(\overline{T}\) is the climatological monthly mean temperature. Selecting \(H\) below the base of the mixed layer allows us to neglect the effect of subgrid-scale processes, such as wind-driven mixing and entrainment, and sunlight penetration on \(T'\) (see DiNezio and Deser [2014], for further details). The heat budget in (A1) is completed with \(Q'_{\text{atm}}\), the net air-sea heat flux (positive into the ocean). The remaining constants are \(\rho_0\), a reference density of sea water, and \(c_p\), the specific heat of sea water.

The difference in total heat content tendency \(Q'_i\) integrated over depth \(H\) between the two climates is plotted in Figure 7c and corresponds to
Figure A1. Climatological mean monthly Surface Downwelling Clear-Sky Shortwave Radiation during boreal summer/fall (JASO) and boreal winter/spring (FMAM) and in (a, b) piControl, as well as the change in the (c, d) mid-Holocene. Zonal deviations are due to the presence of water vapor.

$$\Delta Q'_w = -\rho_0 c_p \left[ \int_{-H}^{0} \frac{\partial T'_{MH}}{\partial t} dz - \int_{-H}^{0} \frac{\partial T'_{piControl}}{\partial t} dz \right] = \sum (A2)$$

$$= -\rho_0 c_p \int_{-H}^{0} \frac{\partial T'_{MH}}{\partial t} dz - \rho_0 c_p \int_{-H}^{0} \frac{\partial \Delta T'}{\partial t} dz = \sum (A3)$$

As discussed in sections 3 and 4, there is a change in the seasonality of SSTs in the tropical Pacific induced by the orbital forcing in the mid-Holocene, as well as a change in the seasonality of the peak of EP events (Figure 4). Since the composite plots presented in section 5 are with respect to the peak of the events and refer to anomalies with respect to each simulation's climatology, the difference plots are indeed descriptive of the differences in the evolution of events within their respective SST (seasonal) background. However, since the two systems (ENSO and background climatology) are not entirely independent, the change in the heating feedbacks may include convoluted changes in ENSO and background climatology that cannot be disentangled by the present analysis.

To illustrate this in detail, consider the difference in upwelling feedback $\Delta Q'_w = -\rho_0 c_p \int_{-H}^{0} \frac{w' \partial T}{\partial z} dz$ that is presented in Figure 7f. The mid-Holocene $w'_{MH}$ can be written as $(w' + \Delta w')$, and the mid-Holocene mean climatological temperature $T'_{MH}$ can be written as $(T' + \Delta T')$. Then, $\Delta Q'_w$ can be written as

$$\Delta Q'_w = -\rho_0 c_p \int_{-H}^{0} \left[ w' \left( \frac{\partial T}{\partial z} + \frac{\partial \Delta T}{\partial z} \right) \right] dz = \sum (A4)$$

$$= -\rho_0 c_p \int_{-H}^{0} \left[ w' \left( \frac{\partial T}{\partial z} + \frac{\partial \Delta T}{\partial z} \right) + \Delta w' \left( \frac{\partial T}{\partial z} + \frac{\partial \Delta T}{\partial z} \right) + \Delta w' \frac{\partial \Delta T}{\partial z} - w' \frac{\partial \Delta T}{\partial z} \right] dz = \sum (A5)$$

The first and last terms in the right-hand side cancel out, and the fourth term can be neglected because it is the product of two $O(\Delta)$ quantities. The third term has the same sign as a potential “independent” ENSO change, therefore cannot be separated. Furthermore, we can ignore this term because ENSO anomalies in the NINO1+2 region are not strictly coupled; rather, they are remotely driven from the central Pacific where ocean-atmosphere coupling is the strongest. And since ENSO SSTs do not change considerably in the central Pacific, we can ignore the $\Delta w' \cdot [...]$ terms of the heat budget.

We are then left with

$$\Delta Q'_w \approx -\rho_0 c_p \int_{-H}^{0} w' \frac{\partial \Delta T}{\partial z} dz \sum (A6)$$
We therefore argue that this term, i.e., the change in upwelling feedback that is due to the change in the climatological stratification, $\Delta aT/\Delta z$, is dominant in producing the reduction in the upwelling feedback shown in Figure 7f and is hence the main cause of difference in the development of ENSO events in the mid-Holocene.

Acknowledgments
This research is funded by U.S. NSF grant OCN-1304910 and U.S. Department of Energy grant DES0005110. We are thankful to the three anonymous reviewers and Associate Editor David Lea for their constructive feedback. Many thanks are due to Bette Otto-Bliesner of NCAR for feedback and support. We wish to acknowledge members of NCAR’s Climate Modeling Section, CESM Software Engineering Group (CSEG), and Computation and Information Systems Laboratory (CISL) for their contributions to the development of CESM and CCSM. All model results are publicly available through NCAR’s portals, and all computed data are available upon request.

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