Eocene Oceanic Responses to Orbital Forcing on Precessional Time Scales

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Abstract. The goal of our study was to gain an estimate of the variability of ocean-related climate processes driven by insolation forcing over a realistic precessional cycle in an Eocene greenhouse world. Between endmembers of a precessional cycle mean annual sea surface temperatures (SSTs) vary by up to 5°C at high northern latitudes, with minimal tropical SST response. Extratropical regions of the Pacific, Atlantic, and Tethys Oceans show up to a two-fold variation in upwelling strength, while oceanic regions adjacent to northwestern Africa, India, and South America exhibit little oceanic upwelling variability. The response of ocean surface moisture balance to the forcing is greatest in the tropics, varying by as much as 60%. Continental runoff varies by up to a factor of two in some regions. These results may be useful in identifying locations with maximum likelihood of future recovery of orbital cyclicity in deep-sea sediments.

1. Introduction

Over the past several years, pre-Quaternary deep-sea sediment records containing orbital-scale variability have been recovered in increasing numbers. Ocean sediment records from the late Cretaceous and Paleocene have suggested the influence of precessional insolation forcing upon deep-sea sediments [Herbert and D'Hondt, 1990; Park et al., 1993; Herbert, 1997]. Sediments of Oligocene to Miocene age have been shown to contain ~40 kyr variability, attributed to obliquity forcing, as well as precession and eccentricity frequency variability [Zachos et al., 1997; Shackleton et al., 1999]. Recently discovered middle Eocene marine sediments from the western Atlantic contain precession-length cycles that may be related to upwelling processes [Kroon et al., 1999; Wade and Kroon, 1999]. Given the continuing advances in scientific deep-sea drilling techniques, there will likely be more such records recovered in the future.

Paleoclimate modeling studies have addressed, to some extent, linkages between orbitally forced insolation variation and atmospheric and oceanic responses in Earth history [e.g., Pbl and Kutzbach, 1992; Crowley et al., 1992, 1993; Park and Oglesby, 1990, 1991; Valdes et al., 1994]. However, there are two limitations found in most modeling studies of pre-Quaternary climates and orbital forcing. First, most studies have not used realistic combinations of orbital parameters. The commonly applied “hot summer orbit/cold summer orbit approach” [e.g., Crowley et al., 1993; Sloan and Morrill, 1998] uses extreme orbital values of precession (the relationship between the solstices and aphelion and perihelion, the positions of largest and smallest distance, respectively, between the Earth and Sun), eccentricity (the degree of ellipticity of Earth’s orbit around the Sun), and obliquity (the degree of tilt of Earth’s rotational axis with respect to the plane of Earth’s orbit). These extreme values occur infrequently in geologic time, according to calculated values of eccentricity, precession, and obliquity for the past 5-10 million years [Berger, 1978]. The second common limitation is that such modeling studies have not included calculated sea surface temperatures (SSTs), which constrains the relationship of model results to understanding linkages between climate and oceanic processes of sedimentation [e.g., Park and Oglesby, 1990, 1991]. Despite these shortcomings, such studies have provided insight into the general relations between insolation variations and climatic responses.

To the best of our knowledge, there is only one published study which used realistic orbital values and calculated ocean surface temperatures [Valdes et al., 1994]. That study focused on the response of Late Jurassic climate to eccentricity-forced insolation changes. On the basis of the results, Valdes et al. [1994] suggested that similar lithologic sequences can represent sedimentary responses to different climate signals in different regions. This conclusion highlights the need to explore orbital forcing in a realistic approach as possible, for specific time periods of interest.

In this study we examine climate variability of the ice-free early Eocene as driven by variations in orbital forcing. A precessional signal has been suggested in several different Paleogene records [Fischer and Roberts, 1991; Roehe, 1993; Kroon et al., 1999], and so we focus on realistic precessional forcing in this climate model sensitivity study. The goal of our study is to gain an estimate of the variability of ocean-related climate processes driven by realistic precessional forcing.

2. Description of Model and Experiments

2.1 Climate Model Description

The climate model used is this study is version 2 of the GENESIS general circulation model (GCM) [Thompson and Pollard, 1997]. This is a spectral atmospheric GCM coupled to submodels of the land surface (including vegetation, soil, and snow systems), sea ice, and a 50 m deep mixed layer ocean. GENESIS has been used extensively for paleoclimate
modeling studies [e.g., Barron et al., 1995; Otto-Bliesner and Upchurch, 1997; Kutzbach and Liu, 1997; Sloan and Pollard, 1998], and the model performs as well as other GCMs in the simulation of present-day climate in terms of capturing large-scale features of the climate system (see more complete description by Thompson and Pollard [1997]).

The surface resolution of the model is 2° latitude by 2° longitude, and the horizontal resolution of the atmosphere is ~3.75° latitude by 3.75° longitude (spectral resolution T31). The 18 atmospheric levels are distributed unevenly in the vertical, with greatest density of levels near the Earth’s surface. The model contains a full seasonal solar cycle and a diurnal cycle. Individual greenhouse gas concentrations are prescribed in the model for CO₂, CH₄, and N₂O. Meridional oceanic heat transport is calculated via a diffusion coefficient that depends on latitude and the zonal fraction of land and ocean at each latitude. Heat flow under sea ice is set to 2 W m⁻² in the Northern Hemisphere and 10 W m⁻² in the Southern Hemisphere. The Norwegian Sea oceanic heat flux used by Thompson and Pollard [1997] and discussed in some paleoclimate modeling studies was not used in our experiments, since it was designed for the present land/sea configuration and climate.

2.2 Eocene Boundary Conditions

Boundary conditions in our study were defined to represent early Eocene physical conditions. The paleogeography, terrestrial elevations, and vegetation are those of Sloan and Rea [1995], interpolated to 2° x 2° resolution. Land surfaces were specified as ice-free. Soil composition was defined with a single average value over land surfaces for all soil layers (set to the present-day average soil composition of 43% sand, 39% silt, and 18% clay) because the global soil composition of the Eocene Earth is unknown. On the basis of limited proxy data and modeling (see summary by Sloan and Rea [1995] and also Pearson and Palmer [1999]), atmospheric CO₂ was set at 560 ppm. Concentrations of atmospheric CH₄ and N₂O were set at preindustrial levels (0.700 and 0.285 ppm, respectively). The atmospheric ozone distribution for the Eocene is unknown, and so this quantity was prescribed at modern values. The solar constant was specified at a modern value (1365 W m⁻²), on the basis of estimated solar luminosity history [Crowley and Baum, 1991].

2.3 Description of Model Experiments

Two experiments were carried out for this study. They had identical boundary conditions, with one exception. The only difference between the experiments was the specified orbital parameters, which were chosen to represent two instants in time from a precessional half cycle. Values for precession, obliquity, and eccentricity were taken from a timeseries of those values calculated for the past 5 million years [Berger, 1978]. The first of the experiments captures a precessional cycle at a minimum value, with precession having a value of 90° (measured as the prograde angle from perihelion to the vernal equinox) and eccentricity and obliquity having values of 0.0531 and 22.8°, respectively. We refer to this as our “MINS” experiment. The MINS orbital configuration produces Northern Hemisphere perihelion during Northern Hemisphere winter solstice and aphelion during summer solstice, resulting in a relatively reduced Northern Hemisphere seasonal insolation cycle. The second experiment represents the precessional cycle ~11,500 years later, when precession is at a relative maximum value (270°) and the other two orbital parameters are altered accordingly (eccentricity is 0.0524, obliquity is 23.8°). We refer to this as our “MAXS” experiment. The MAXS orbital configuration results in Northern Hemisphere perihelion during summer solstice and aphelion during winter solstice. These conditions produce an amplified seasonal cycle of solar insolation for the Northern Hemisphere. The prescribed forcing results in the largest insolation difference at high latitudes between the cases (Figure 1).

This study is somewhat unique because we have chosen to vary eccentricity and obliquity along with the precession component of the orbital forcing rather than carrying out a more standard climate model sensitivity study where only one factor (such as precession) is altered between two model experiments (however, an analogue may be found in climate modeling studies of the late Quaternary); [e.g., Cooperative Holocene Mapping Project (COHMAP) Members, 1988]. Our rationale for this approach was to employ as realistic a forcing (i.e., a change from one combination of precession plus eccentricity plus obliquity elements to another combination) as possible because we wanted to gain an estimate of the variability of climate processes that would occur across a realistic precessional cycle.

Given our goal, in the experiments, precession varies in conjunction with eccentricity and obliquity, just as these components have been calculated to vary [e.g., Berger, 1978]. As a result, the small changes in obliquity and eccentricity (1.0° and 0.0007, respectively) that occur between the two experiments contribute to the model results. In the remainder of the manuscript, when we use the terms “precession-driven forcing” and “prescribed orbital forcing,” they are meant to include the total insolation change caused by the change in all of the orbital parameters over our precessional cycle. However, we have determined that the differences in eccentricity and obliquity between the cases causes <5 W m⁻² (concentrated at midlatitudes) between the cases on a mean annual basis.

![Figure 1](image-url)
3. Results

In this paper we report only the surface oceanic response (as coupled to the atmosphere) in order to focus on model results and processes that may be related to deep-sea sediment records. The terrestrial climate responses will be reported in a later paper (K. Lawrence et al., manuscript in preparation, 2000).

3.1 Sea Surface Temperatures and Sea Ice

SSTs show substantial response (i.e., large sensitivity) to the precessional-driven forcing (Figure 2). The greatest SST response is seen at high latitudes owing to a combination of the extreme insolation differences at these latitudes (Figure 1) and the resulting differences in sea ice extents (Figure 3). On an annual basis, SSTs vary by up to 6°C in northern ocean regions between the MAXS and MINS results (Figure 2). Equatorward of ~60° latitude, there is very little difference in mean annual SST values.

For the period of December through February (DJF), SSTs in the Arctic and North Pacific Oceans vary by up to 10°C, with the MINS results containing lower-temperature values (Figure 2). The variability of response is due partly to direct heating consequences of the altered incoming solar radiation but more importantly to the indirect effects of sea ice. The MINS experiment contains ~20 W m⁻² more winter insolation at 80°N, but this small insolation difference results in a small temperature response. More important is the summer heating, which is ~120 W m⁻² less in the MINS case at 80°N (Figure 1), and the resulting sea ice. The lower DJF temperatures in the MINS results are caused by the more extensive summer sea ice in that case (Figure 3), which prevents the high-latitude ocean from warming during the summer and provides an easy cooling in the fall and winter. At high southern latitudes, SSTs differ by 2°-3°C between DJF results, and SSTs are nearly the same everywhere else in both experiments.

For June through August (JJA), SSTs differ by at least 2°C poleward of ~40° latitude in the Northern Hemisphere, driven by the large difference in incoming solar radiation (up to 100 W m⁻² at 45°N and up to 120 W m⁻² at 75°N) between the experiments (Figure 1). Much of the Tethys, North Pacific, and North Atlantic ocean surface is 2°-3°C warmer in the MAXS results. As discussed above, there is a substantial difference in Arctic sea ice distributions between the cases in JJA. There is greater Arctic sea ice extent in the MINS case, caused by the reduced summer solar insolation in that case. In the Southern Hemisphere, at high latitudes, SSTs vary by 1°-3°C, with higher values in the MINS results. That the high-latitude winter SST response is smaller in the Southern Hemisphere than in the Northern Hemisphere can be explained by the solar insolation distribution; the prescribed orbital forcing has a greater impact on Northern Hemisphere DJF and JJA periods than on the same periods (especially JJA) in the Southern Hemisphere (Figure 1).

3.2 Sea Level Pressure

Variations in DJF and JJA sea level pressure show substantial differences between the cases in both seasons and hemispheres in response to the precessional-driven insolation changes. In DJF the Aleutian and Icelandic Lows and the Siberian High are strengthened in the MINS results relative to the MAXS results (Figure 4). There is greater Northern Hemisphere insolation in the MINS case at this time (DJF), which works to enhance the lows via intensified surface heat-
ing. The high-latitude Southern Hemisphere low-pressure systems and subtropical high are increased in the MINS case during DJF for the same reasons. In JJA the Central Asian Low and the Pacific High are much stronger in the MAXS results (Figure 4) because of enhanced insolation in those regions (Figure 1). In the Southern Hemisphere the high-latitude lows are slightly weaker in the MAXS results.

3.3 Surface Winds

The sea level pressure differences lead to numerous differences in surface wind patterns between results of the two experiments. While the overall variability in wind speed and direction that occurs in response to the orbital forcing might not seem large (Figure 5), these differences set up substantial variations in wind-driven upwelling (below). In DJF winds over the North Atlantic are stronger in the MINS results (by up to 4 m s\(^{-1}\), a 30\% change). Surface winds over the North Pacific are stronger in the MINS results (by 2-3 m s\(^{-1}\), a 30\% change). There are only minor differences in surface winds between the cases in the Southern Hemisphere during DJF; most notably, there are slightly stronger winds alongshore of Antarctica in the MINS results.

In JJA, greater insolation and stronger Northern Hemisphere heating in the MAXS experiment produces stronger surface winds in that case (Figure 5). There are relatively strong alongshore winds out of the south adjacent to Asia (up to 9 m s\(^{-1}\) in the eastern Pacific in the MAXS results, while the MINS results in the same region have winds that are very weak (maximum velocities of ~5 m s\(^{-1}\)) and variable in direction (Figure 5). Surface winds over northeastern Tethys are nearly twice as strong in the MAXS results relative to the MINS results. In addition, wind directions in this region are highly variable between the cases.

3.4 Wind-Driven Upwelling

We calculated the wind-driven upwelling that model winds would create if our model included a dynamic ocean component. In general, upwelling is driven by wind stress curl. The relationship between winds and upwelling can be expressed as the divergence \(\nabla \cdot \mathbf{v}_E\) of the Ekman transport \(\mathbf{v}_E\),

\[
\nabla \cdot \mathbf{v}_E = \frac{-1}{\rho_w} \mathbf{k} \cdot \nabla \times \left( \frac{f}{F} \right),
\]

(1)

calculated from curl \((\mathbf{k} \cdot \nabla \times \mathbf{v})\) of the wind stress, \(\mathbf{v}\), (calculated from AGCM lowest atmospheric level wind fields) [Trenberth et al., 1990]. In this expression, \(f\) is the Coriolis parameter, and \(\rho_w\) is the density of seawater (1027 kg m\(^{-3}\)). The Ekman transport divergence provides a reliable estimate of wind-driven upwelling rates in the present day [Hellerman and Rosenstein, 1983; Pedlosky, 1996], and GENESIS present-day
model results compare well with observations (M. Huber and L. C. Sloan, manuscript in preparation, 2000).

The calculated wind-driven upwelling shows a strong sensitivity (i.e., large variability) to the insolation forcing in our experiments, with upwelling strength varying by a factor of 2 between cases. Overall, stronger and more widespread upwelling occurs in the MINS results in both seasons (Plate 1). Owing to the stronger surface pressure highs and lows in the MINS results, DJF upwelling in the northwestern Pacific, northern Atlantic, and northern Tethys is much stronger and more widespread (by up to 2.5 cm s\(^{-1}\), a near doubling in upwelling rates) relative to the MAXS results (Plate 1). Similarly, there is increased and more widespread upwelling in the MINS case in the Southern Ocean and along the northern shore of Australia (by up to 1.25 cm s\(^{-1}\)), associated with the stronger lows in those regions. Upwelling shows little variability and is vigorous in both cases in the regions surrounding India, along portions of the western margin of Africa, and along the western and northern margins of South America (Plate 1).

In JJA, upwelling is slightly stronger in the MAXS case (by ~1 cm s\(^{-1}\)) along the western margin of the North Atlantic and in the northern Tethys in limited regions (Plate 1). In association with the stronger lows in the JJA MINS results, upwelling is nearly twice as strong and more widespread in the Southern Ocean (relative to the MAXS results). In both experiments, there is little upwelling variability in regions along the northern and western coasts of South America and the western coast of Africa and in the region between Africa and India; upwelling is consistently strong here with both forcing combinations.

3.5 Precipitation Minus Evaporation (Surface Moisture Balance)

The net moisture balance at the ocean surface is calculated here as the difference between surface precipitation and evaporation. The balance between precipitation and evaporation is largely determined by surface temperature (both precipitation and evaporation respond nonlinearly to increasing temperature) and by wind speed (evaporation increases with increasing surface wind velocity). For both DJF and JJA the largest variability in net surface moisture (up to 3 mm/day, a 60% change) in response to the prescribed insolation forcing is found near the equator (Figure 6). Since this is the region of the globe with the highest temperatures, the balance between precipitation and evaporation shows the greatest variability in this region. This response reflects the difference in received insolation between the two cases (Figure 1) and the associated strength of the intertropical
convergence zone (ITCZ) that is driven by the tropical insolation and heating. In the extratropics, variation in net surface moisture balance between the MINS and MAXS results is greatest at high northern latitudes in both seasons (Figure 6). In JJA, there is up to 0.5 mm day$^{-1}$ greater (a 50% change) (more positive) net surface moisture balance in the MAXS results. In these latitudes during JJA, there is more insolation in the MAXS case (Figure 1), which produces relatively warmer air and more precipitation. Evaporation in these regions does not increase by the same amount because temperatures are relatively low compared to tropical values.

3.6 Continental Runoff

Continental runoff in the climate model used in this study (like most climate models) is a relatively simple parameterization of continental hydrologic processes (see description by Thompson and Pollard [1997]). Runoff is calculated in the model on the basis of a balance of precipitation and ground saturation, but this volume does not return to the model ocean in a realistic distribution (also note that there is no deep ocean component to this model). However, we consider runoff results here because continental runoff can affect many processes that can be recorded by deep-sea sediments [e.g., Dean and Gardner, 1986; Raymo et al., 1988; Froelich et al., 1992; Sloan et al., 1997].

On the basis of our paleogeography and paleotopography boundary conditions we have identified continental-scale drainage basins for our model world, and we examine the annual amount of continental runoff that would be shed to the oceans from these basins if runoff was more realistically coupled to the oceans. We calculated the total annual continental runoff from each basin per year and the difference in total annual runoff between the MINS and MAXS results (Figure 7). The results do not show any broad trends, with the exception of greater runoff (by up to 42%) in the MINS results for all of Australia and Antarctica. In those regions the orbital forcing produces greater snow accumulations (not shown) and greater runoff in the MINS experiment. The opposite effects occur in the MAXS results in the Northern Hemisphere. There is an increase in runoff due to snowmelt (not shown) for the higher-latitude and higher-elevation drainage basins of North America (by up to 22%) but not over the lower-elevation northern Eurasian basins. In these latter areas, there is little snow accumulation and little runoff as a result. The largest variability in continental runoff between the MINS and MAXS experiments occurs in areas dominated by tropical monsoon
Plate 1. Wind-driven upwelling for (a) DJF MINS case, (b) DJF MAXS case, (c) JJA MINS case, and (d) JJA MAXS case. Green to red colors indicate upwelling from weak to strong intensity. Negative values (blues to white) indicate downwelling. Hatched regions indicate where upwelling calculations are invalid.
activity, northern Africa and northern South America (Figure 7). In these regions the annual total runoff to the tropical Atlantic varies by as much as 53% between the MINS and MAXS results and by up to 106% into southern Tethys from north Africa.

4. Discussion

Our goal in this study was to gain an estimate of the variability of ocean-related climate processes as driven by realistic precessional forcing in an Eocene greenhouse world. Ideally, in the future, this approach would be taken with many more modeling experiments, in order to create more fully defined, orbital-scale variability estimates of processes that may be related to marine sedimentation. We have examined only two settings of orbital parameters and one rendition of early Eocene boundary conditions, and yet the results are instructive in that they provide one estimate of variability of climate processes related to realistic orbital forcing. We address this issue in section 4.1. The results also provide a basis for comparison to deep-sea sediment records. We make a limited comparison to such records in section 4.2.

4.1 Variability of Climate Processes Influenced by Precessional Forcing in an Eocene World

Our results show that changes in insolation over a realistic precessional cycle can drive a large amount of ocean-related climate variability during the Eocene greenhouse interval. This variability is manifested primarily as changes in surface temperature, upwelling intensity and location, net surface moisture balance, and amount of continental runoff. Temperature and upwelling responses to the forcing are greatest in regions outside of the tropics (Figure 2 and Plate 1), while the surface moisture response is largest in the tropics (Figure 6). Continental runoff is highly variable across all latitudes in response to the prescribed orbital forcing (Figure 7). Overall, the precession-driven variability of these climate processes is of large enough magnitude that it may be expressed in the geologic record in many regions, given proper sedimentation processes, rates, and preservation.

Mean annual SSTs at latitudes poleward of $60^\circ$ show variability of up to $5^\circ$C in response to the precessional forcing; there is very little mean annual temperature response in the tropics (Figure 2). The muted response of tropical SSTs is due to the smaller mean annual variation in insolation between the cases at those latitudes (Figure 1) and is also due to the high heat capacity of water (there is a much larger temperature response on land) (Figure 2 and see also K. Lawrence et al., manuscript in preparation, 2000).

Winter SSTs show a response twice as large as the mean annual response, concentrated at high northern latitudes with our prescribed forcing. This response is primarily a function of large insolation variations at high latitudes (up to 120 W m$^{-2}$ between cases; see Figure 1) and of sea ice differences between the cases. The sea ice responses to forcing represent a strong seasonal lag factor in the high-latitude climate system. In a greenhouse climate, sea ice may drive high-latitude temperature responses to orbital forcing. Additionally, the high-latitude response of winter cooling may influence deep ocean convection in these regions. However, this last point is only speculation since our model lacks a full ocean component.

The summer temperature response to the orbital forcing is smaller (~2°-3°C) than the winter response, but it is more widespread, especially for the Northern Hemisphere. This results suggest that the seasonal signature and amplitude of temperature change in response to the forcing may be hidden in the mean annual record. Summer temperature conditions should correlate best to growing seasons for marine flora and fauna, and as a result, summer SST values may have a greater influence on paleo-SST interpretations than mean annual values. If so, our results suggest that orbital-scale temperature variability interpreted from marine sediments may more reflect orbitally modulated changes in seasonal temperatures than mean annual temperatures.

Wind-driven upwelling exhibits large sensitivity to the imposed orbital forcing in all oceans. In general, meridional pressure gradients are stronger in the MINS case in both hemispheres during both seasons (Figure 4), with the exception of the MAXS case in JJA (due to the increased summer insolation in that case). As a result, upwelling intensity is
increased (in some cases by a factor of 2) in both hemispheres in specific regions (e.g., North Pacific, North Atlantic, and Southern Ocean) in the MINS results (Plate 1). There is a seasonal component to the upwelling sensitivity in many of these regions (for example, off the east coasts of Asia and North America), especially at midlatitudes. In contrast, tropical ocean regions adjacent to northwestern Africa and South America show very low upwelling sensitivity to changes in orbital forcing. If we use upwelling as a proxy for marine productivity, our results indicate that productivity intensity is strengthened and weakened over a precessional cycle in many regions (e.g., along the eastern coasts of Eurasia and North America and in the polar front region of the Southern Ocean).

The net surface ocean moisture balance shows sensitivity to the orbitally forced insolation changes at low latitudes despite the small temperature response in these latitudes. These net moisture changes in these areas are due to changes in the solar insolation distribution, which in turn, drive differential heating and control the strength and location of the ITCZ. In the tropics the MINS case has up to 70 W m\(^{-2}\) more solar insolation in DJF, while the MAXS case has 60-80 W m\(^{-2}\) more insolation in JJA (Figure 1). This insolation difference translates into a net precipitation difference, while evaporation is similar in both cases in the tropics. There is also a noticeable difference in the surface moisture balance between cases at high northern latitudes. Such variability may drive or suppress localized vertical ocean mixing and deep convection.

There is large variability of continental runoff in response to the imposed insolation forcing (Figure 7). Runoff from continental-scale drainage basins changed by over 40% between the MINS and MAXS results in several regions. There are numerous possible implications for marine sedimentation processes. Increased runoff could be associated with increased clastic sediment input and possible carbonate dilution, greater freshwater flux to the surface ocean and perturbation of vertical ocean mixing regimes, and/or increased input of nutrients from continental sources to the surface ocean. Without a more detailed model that includes a fully resolved ocean, sediments, and biogeochemistry, we cannot determine the true impacts of our runoff results on marine sedimentation regimes. While the implications for the sediment record are not clear, the magnitude of runoff variability suggests that runoff would have been important in determining the nature of marine sedimentation and that it would be unwise to assume similar runoff patterns from different regions as climate changes, even on orbital timescales.

4.2 Comparison of Model Results to Deep-Sea Sediment Records

Many studies have identified or hypothesized links between cyclic lithologic variations and climate forcing. Lithologic variations have been extensively linked to processes that include carbonate production, clastic input/dilution, nutrient input, ocean circulation characteristics, sea level variation, and oxidation/reduction conditions [e.g., Dean and Gardner, 1986; Bottjer et al., 1986; Roof et al., 1991]. However, we cannot evaluate the role of these processes because they are not included in our modeling framework. What climate modeling can provide is an indication of where some of these mechanisms might act, from a climate perspective.

While it is important to compare model results and paleoceanographic data in order to evaluate model performance, evaluate hypothesized causes of climate change, or assess our
interpretations of proxy climate data, we should also consider the validity and limitations of such comparisons. There are three broad limitations to the comparisons between geologic data and model results that we can make. First, there is a time constraint. While our model boundary conditions represent the early Eocene, they might, by some interpretations, as easily represent the late Paleocene or late early Eocene, given the model spatial resolution and the specified boundary conditions. Also, even if we were confident that our sensitivity study was a representative simulation, we would have to restrict the temporal resolution of the proxy climate data that we use for comparison to our model results in order to represent time as precisely as possible. If we limit the data that we use for comparison to (for example) a 3 million year time range, we are encompassing a large range of orbital forcing in terms of precession, obliquity, and eccentricity values, and we have only carried out two such combinations of values in this study. Thus we have potential offset or mismatch built into our model results and geologic proxy climate data from a temporal perspective.

Second, there is a spatial limitation to all comparisons of model results and proxy climate data. The surface resolution of our model is 2° latitude by 2° longitude, which means that a grid cell near the tropics is ~49,454 km². A climate model will produce a single value (of temperature, upwelling, etc.) for a grid cell of that area, while the variation in character of real climate will have much greater variability on smaller spatial scales. Thus the model is not capturing subgrid-scale features of climate processes that may be recorded by proxy climate data. This spatial difference also introduces a fundamental discrepancy between the model results and the proxy data interpretations.

As a third limitation to the comparisons, we must consider the processes that are included (and omitted) from a climate model world versus those of the real world. Any differences between these "worlds" may have an impact on the comparison results and prohibit a perfect correlation between model results and proxy data interpretations. We note three key differences that have a bearing on the following comparison. (1) Our modeled upwelling is solely driven by wind and has no connection to the deeper ocean. (2) We have no deep-ocean component to our model and so lack the influence of ocean circulation dynamics and bathymetry upon continental shelves and deep ocean currents. (3) Our model lacks any biogeochemical components, including feedbacks to the climate system.

As a result, the best motivation for comparing the model results with geologic data in this study is to investigate the hypothesis that orbital forcing may explain some of the variability seen in climate records during greenhouse climate intervals. We can test this hypothesis by comparing our results to early Eocene marine sediment data to see if the model results show variability in regions containing such data. However, as outlined above, the results of such comparisons are only instructive and are not conclusive.

There are currently few geologic records to test our findings. There is evidence for precessionally forced sediment cyclicity in the western Atlantic (Blake Nose) of middle Eocene age [Kroon et al., 1999; Wade and Kroon, 1999] and in the southern Atlantic (~30°S paleolatitude) for the early Paleocene [Herbert and D'Hondt, 1990]. We have identified different combinations of climate processes that vary in those regions. Our results of relatively large temperature and upwelling and runoff responses (Figures 2 and 7 and Plate 1) are consistent with the sediment records from the western Atlantic, and our runoff results (Figure 7) are consistent with findings from the southern Atlantic in that the model results identify processes that show substantial variability in response to precessional forcing.

The western Atlantic records [Kroon et al., 1999] show precession-length cycllicity in a region where our results indicate large upwelling sensitivity to precessional forcing (Plate 1). Upwelling intensity is greatest in winter months, but there is significant variability between the MINS and MAXS results in both DJF and JJA. The results suggest that cyclic sedimentary sequences in this area may be due to sedimentation processes associated with upwelling (e.g., productivity). Our results also show a 32% change in the amount of continental runoff between the cases. Thus runoff may have also influenced the sedimentary system in the Blake Nose region and may explain some of the lithologic cycllicity in that area. Last, summer SSTs show a 2°-3°C variation in response to the precessional forcing. This temperature difference also may have contributed to the sedimentation regime via biogeochemical feedbacks.

At the southern Atlantic site of Herbert and D'Hondt [1990] the model results show a response to orbital forcing only in the form of continental runoff. Runoff in the subtropical south Atlantic varies from +6 to -7% between the cases. Whether or not this is sufficient to explain the variations in carbonate content documented by Herbert and D'Hondt [1990] and Herbert [1997] is unknown. If the variation in continental runoff does not explain the sedimentary record, the lack of correlation between our results and the sedimentary record in this region could be due to any of the limitations discussed above.

Our results also predict where additional precessional-frequency cyclic sedimentary records might be found (but with the same three general limitations discussed above). We predict that the northwestern Pacific, northwestern Atlantic, northeastern Tethys, and the Southern Ocean adjacent to northeastern Australia should contain records of Eocene precessional insolation forcing if sedimentation in those regions is linked to temperature and/or upwelling conditions (Figure 2 and Plate 1), and if there is sufficiently high sedimentation rate and adequate preservation. Changes in runoff are greatest in the Southern Ocean, central Tethys, and the central and northern Atlantic Ocean; these regions may yield cyclic sedimentary records as well, with the same caveats as above. The results also suggest that orbital cyclicity seen in SSTs at midlatitudes may reflect changes in SST seasonality more than changes in mean annual temperature values. Precessional variability in the tropics may be most strongly manifested as oceanic processes associated with continental runoff and the net moisture balance at the ocean surface (Figure 6). The modeling approach applied here may be used to predict areas where orbital forcing should generate high amplitude and easily tuned records for the construction of astronomical timescales [e.g., Zachos et al., 1997; Shackleton et al., 1999; Herbert, 1999].
5. Conclusions

Our results show that there is significant ocean process variability driven by orbital forcing in an Eocene greenhouse world. The response is seen prominently in SSTs and wind-driven upwelling in the extratropics, surface ocean moisture balance in the tropics and at high northern latitudes, and continental runoff worldwide. Our results predict that orbitally forced upwelling processes should have influenced sedimentation in regions of all of the Eocene oceans.

Modeling studies such as this one may be of more applied use in the acquisition and interpretation of deep-sea sediments than previously realized. Our study demonstrates the capability of using climate model results to identify regions of high sensitivity to forcing via definition of processes tied to those recorded by deep-sea sediments.

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