A Role for the Tropical Pacific Coupled Ocean-Atmosphere System on Milankovitch and Millennial Timescales. Part I: A Modeling Study of Tropical Pacific Variability

Amy C. Clement [1] and Mark Cane

Lamont-Doherty Earth Observatory of Columbia University, Palisades, NY

Climate records from over much of the world show variability on both Milankovitch and millennial timescales. However, mechanisms in the climate system that have a global scale are lacking. In this two-part paper, we turn attention to a part of the system that is known from the modern climate record to be capable of organizing global scale climate events: the tropics. In the first part, we isolate this system using a coupled ocean-atmosphere model of the tropical Pacific. The model results demonstrate a tropical mechanism which can amplify Milankovitch forcing and generate a mean climate response. The model is also run for 150,000 years with no forcing. The results raise the possibility that millennial timescale variability can be generated within the tropics through non-linear interactions. In that case, power is to be expected over a range of frequencies, rather than in any particular narrow band. The second part of our study, presented in the following paper, will discuss the potential impact on the global climate of these mechanisms of tropical climate variability.

1. INTRODUCTION

The El Nino/Southern Oscillation (ENSO) is perhaps the most well-studied modern climate phenomenon. The observational record indicates that ENSO tends to have most power at 2-7 year periods. At the heart of ENSO is a positive feedback between the ocean and atmosphere. A reduced equatorial sea surface temperature (SST) gradient, as in the warm phase of ENSO, leads to a slackening of the trade winds, reduced equatorial upwelling, and a deepening of the thermocline in the eastern equatorial Pacific, all of which further weaken the SST gradient [*Bjerknes*, 1969; *Cane*, 1986]. In the cold phase, the feedbacks push the system in the opposite direction. An increased equatorial SST gradient is enhanced by stronger trades, more equatorial upwelling, and a steeper thermocline tilt. These interactions are not restricted to the 2-7 year timescale. The modern climate record indicates that ENSO may have variability on decadal timescales [*Trenberth and Hurrell*, 1994]. Whether this variability is internal to the tropics [i.e., *Zebiak and Cane*, 1991], or is the result of extra-tropical forcing of ENSO [i.e., *Kleeman et al.*, 1999], the same positive feedbacks will operate on this longer timescale.

In addition to the potential for natural, perhaps internal low-frequency tropical variability, there is also the great likelihood that ENSO physics play an important

[1] Now at LODYC, Université de Paris 6, Paris, France.

Mechanisms of Global Climate Change at Millennial Time Scales
Geophysical Monograph 112
Copyright 1999 by the American Geophysical Union

role in the response of the climate to an external forcing [*Clement et al.*, [1996]; *Dijkstra and Neelin*, 1995; *Cane et al.*, 1997]. There is also the possibility that the ENSO behavior will be altered by such a forcing. Whether ENSO has changed in response to greenhouse forcing is currently an issue of debate in the climate community [*Trenberth and Hoar*, 1996; *Rajagopalan et al.*, 1997; *Latif et al.*, 1997; *Goddard and Graham*, 1997]. Modeling studies have shown that the temporal characteristics of ENSO may be altered by greenhouse forcing [*Zebiak and Cane*, 1991; *Clement et al.*, 1996; *Timmermann et al.*, 1998].

Separating natural and forced variability embedded in the tropical Pacific in the modern climate record is difficult because the same coupled physics can be instrumental in the response to a forcing and in generating natural low-frequency variability. The result is a spatial pattern of tropical Pacific SST variability in the 20th century that can look remarkably similar over a range of timescales [*IPCC*, 1995; *Zhang et al.*, 1997; *Latif et al.*, 1997]. Is this low-frequency variability of ENSO or a change in the mean state? Is it forced or natural? The simplest answer is that the tropical Pacific climatology must be thought of as the result of similar physics, whether forced or arising from internal instabilities, operating on a variety of timescales that interact with each other, and cannot be separated.

In this paper, we extend the study of ENSO-like variability to Milankovitch and millennial timescales. The approach we take is to isolate the tropical Pacific using a simple coupled ocean-atmosphere model. The focus of this work is on determining to what extent the tropical Pacific climate can change *on its own* with no influence from higher latitudes. We will explore the mechanisms of climate change on these timescales and make some tentative statements about what kind of temporal variability to expect from this system alone.

2. MODELLING EXPERIMENTS

We perform experiments with the Zebiak-Cane ENSO model [*Zebiak and Cane*, 1987]. This is a coupled model of the tropical Pacific which computes anomalies in the circulation and SST about a mean climatology that is specified from observations. The model domain is 124°E to 80°W and 29°N to 29°S. The dynamics in the atmosphere and ocean are described by linear shallow water equations on an equatorial beta-plane. In the ocean, an additional shallow frictional layer of constant depth (50 m) is included to account for the intensification of wind driven currents near the surface. An atmospheric heating anomaly is computed from the SST anomaly, and the specified background wind divergence field. Wind field anomalies are computed from this heating, and are used to drive the ocean model. The anomalous ocean circulation and thermocline depth are used to compute a new SST anomaly. The subsurface temperature anomaly is a non-linear function of the thermocline depth anomaly. This coupled model produces self-sustained interannual oscillations and contains the main physics thought to be relevant for ENSO.

The model is run for the past 150,000 years and forced with variations in solar radiation due to changes in eccentricity, obliquity, and precession of the equinoxes [*Berger*, 1978]. The Milankovitch solar forcing is implemented as an anomalous heat flux into the ocean surface. The solar radiation anomaly relative to today is computed as a function of time in the past (Kyr), season (τ), and latitude (θ), and converted to a surface heat flux by:

$$F_o'(Kyr,\tau,\theta) = Q_s(T_a + 0.0019\theta_z) - F_o(0,\tau,\theta) \quad (1)$$

where $T_a = 0.7$ which is the transmissivity of the atmosphere meant to represent the bulk effect of absorbtion and reflection of solar radiation by all atmosphere constituents, Q_s is the derived solar radiation at the top of the atmosphere, θ_z is the solar zenith angle [*Berger*, 1978], and $F_o(0,\tau,\theta)$ is the modern distribution of surface solar radiation similarly computed. In addition, a 150,000 year control run is performed where there is no external forcing.

2.1. Milankovitch Forcing

Figure 1a shows 500 year averages of the NINO3 index (the SST anomaly averaged over 150°, 90°W and 5°S, 5°N) for the model experiment with Milankovitch forcing. NINO3 is generally taken to be an index of ENSO. The mean NINO3 value for the control run is about $0.4K$ [*Zebiak and Cane*, 1987], and is subtracted from the NINO3 index of the forced run. This 500 year average index has a large precessional peak (about 21 kyr) as well as significant power at about 11 kyr. The 11 kyr cycle will be discussed elsewhere. The obliquity cycle (41 kyr) is not present. Warm periods generally occur when perihelion (anomalous heating) occurs between December and June, and cold periods occur when perihelion occurs between July and November. The spatial pattern that arises on the precessional timescale is shown in Figure 2. This is the first empirical orthogonal function (EOF) which describes 75% of the variance of the lowpass filtered (> 20 kyr periods) SST field. The pattern looks much like ENSO with the largest signal occuring in the eastern equatorial Pacific.

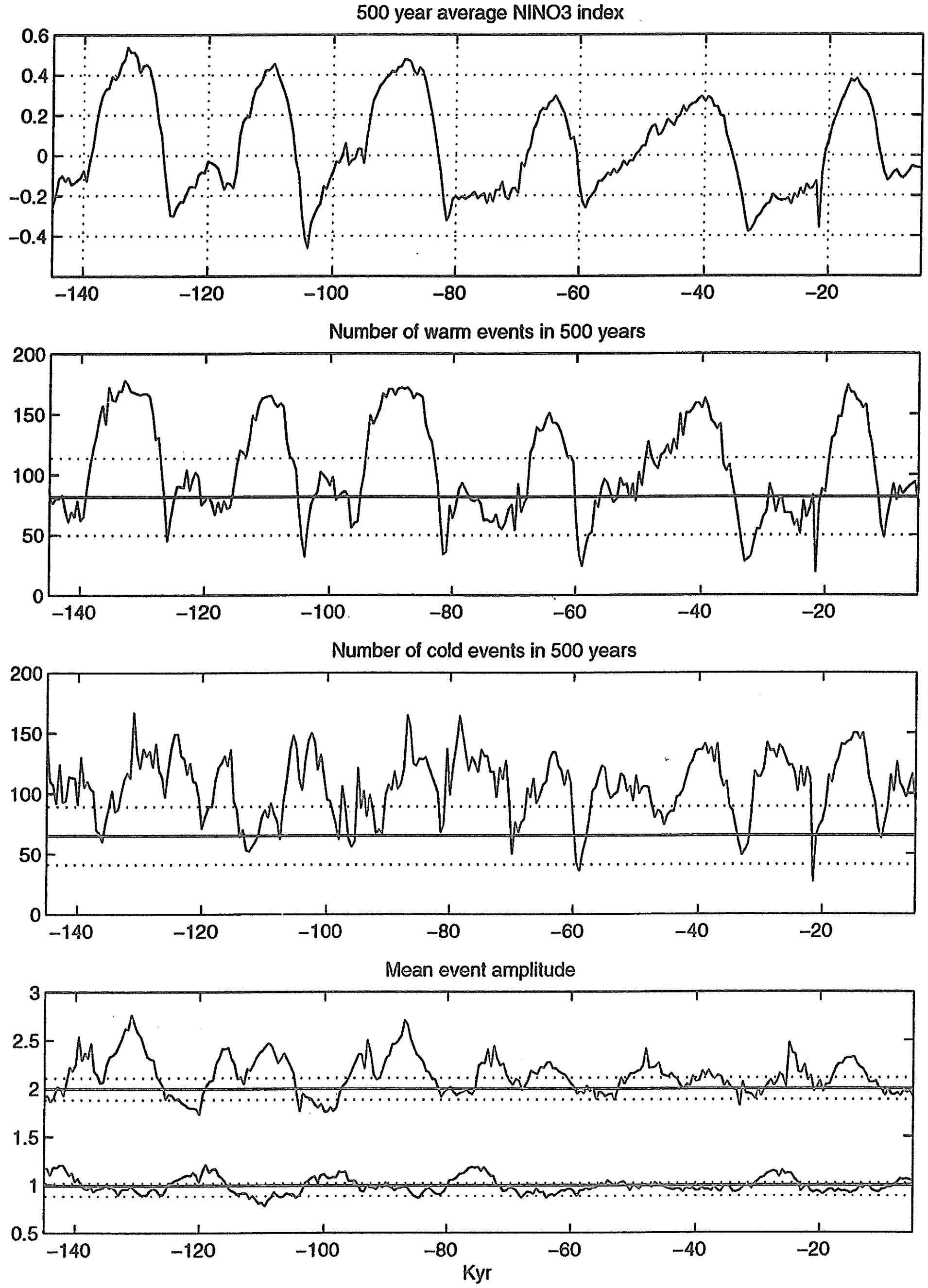

Figure 1. (a) 500 year average NINO3 (degrees Celsius) index from Zebiak-Cane model forced with Milankovitch solar forcing, (b) number of warm events defined as in text for 500 year non-overlapping periods, (c) number of cold events (d) mean amplitude of warm (solid) and cold (dashed) events. The bold line shows the mean values for each of these statistics for the control run with 95% confidence limits plotted (dotted).

What causes this annual mean response to a forcing that has approximately zero annual mean? Figures 1b and 1c show the number of warm and cold events in non-overlapping 500 year windows. A warm event is defined here to occur when the annual mean value of NINO3 exceeds 1 K, and a cold event when the index is less than -1 K. The mean amplitude of events is also shown (Figure 1d). Generally, warm periods occur when

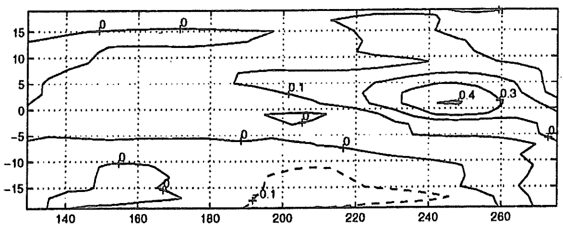

Figure 2. The first EOF (75% of the variance) of SST data with a lowpass filter applied (> 20 kyr) for the Milankovitch forcing run.

there are more frequent and larger warm events while cold periods occur when the cold event frequency and amplitude are larger.

A look at the time series shows more clearly how the character of the events changes. Figure 3 shows 100 year segments from the control run NINO3 time series, from typical warm periods (i.e. 135 kyr), and typical cold periods (i.e. 75 kyr). In the control run, the peak frequency occurs at about 4 years. During the warm periods, events are larger and more regular, and the peak frequency shifts to approximately 3 years. During the cold periods, cold events tend to be more frequent but there is a less well defined peak frequency, and the interannual variability is effectively spread out over a range of frequencies.

While the time evolution of the events change, the spatial structure remains essentially the same. The first EOF of the annual mean SST, thermocline, and wind field during the warm and cold periods are essentially the same as those for the control run (Figure 4). These patterns are described by *Zebiak and Cane* [1991] as the mature ENSO signal. Thus, the orbitally induced changes in the interannual variability result from a change in the time evolution of the events, while the basic coupled dynamics are unaltered. The persistence of the fundamental dynamics explains why the spatial pattern of SST change on a 21 kyr timescale looks similar to ENSO (Figure 2).

The reason for the change in the ENSO variability is discussed in *Clement et al.* [1999] in detail. In brief, it is the result of a seasonal cycle in the response of the system to the forcing. Consider a uniform heating of the tropical Pacific, which approximates the precessional forcing. When a uniform heating is applied, it will initially generate a warm SST anomaly which will affect the atmosphere differently in different seasons. For example, in spring, the ITCZ is near the equator and the wind field is convergent all across the equator. Thus, the uniform SST anomaly will generate a more or less zonally symmetric response in atmospheric heating. However, in late summer/early fall, the eastern Pacific ITCZ moves north, and the wind field becomes divergent in the east, while still convergent in the west. Thus, the warming of the tropical ocean yields a larger heating of the atmosphere in the western Pacific where the mean (background) wind field is already convergent, than in the eastern Pacific, where the strong mean divergence suppresses the development of deep convection. The zonal asymmetry in atmospheric heating anomaly drives easterly wind anomalies at the equator. The coupled system amplifies this perturbation on an interannual timescale via the same unstable interactions that give rise to ENSO, and the result is a cooler east Pacific, or more La Niña-like conditions. A uniform cooling yields the opposite response: little response to the forcing in spring, while in the late summer/early fall the result is *westerlies* on the equator which can develop into a more El Niño-like response. Thus, the mean change in ENSO is dictated primarily by the forcing in the late summer with warming in the summer giving a La Niña-like response and cooling in the summer giving an El Niño-like response.

2.2. Low Frequency Variability in the Control run

The 150,000 year control run is analyzed to assess the low frequency variability of this system with no external forcing. Figure 5 shows the spectrum of the annual mean NINO3 index computed using a multitaper method with non-overlapping windows of length

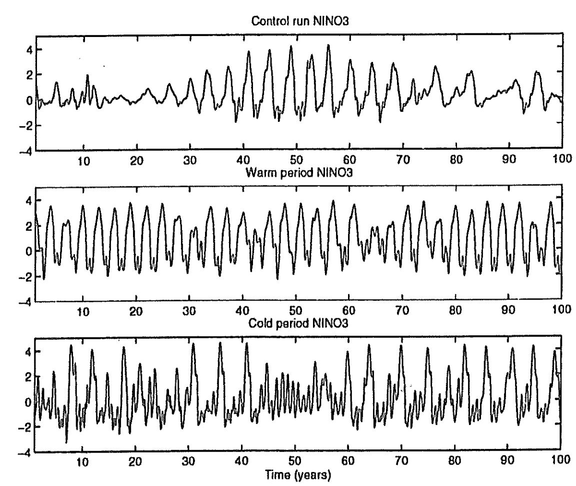

Figure 3. NINO3 time series segments from (a) control run (b) warm period of the forced model run and (c) cold period from the forced model run.

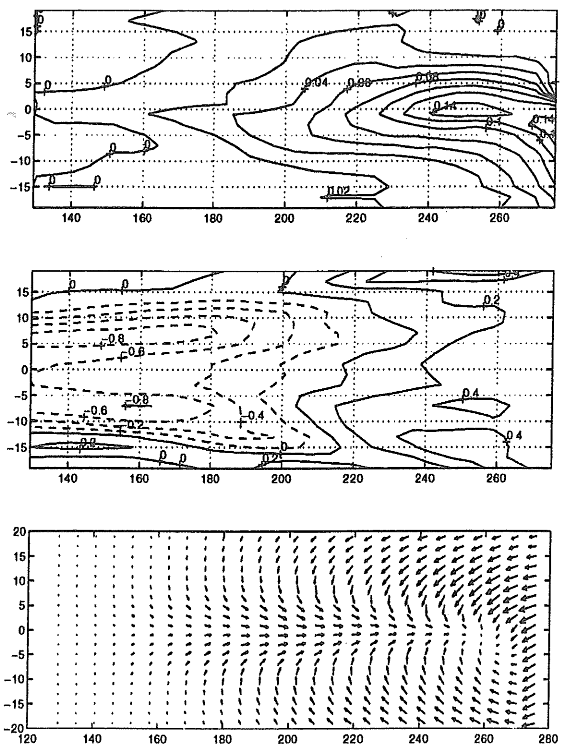

Figure 4. Leading EOF of (a) SST (85% of the variance) (b) thermocline (55% of the variance) and (c) wind field vectors (80% of the variance) for annual mean values from the control run.

2^{14} years [*Mann and Lees*, 1996 and references therein]. This system has a known peak at 4 years [*Zebiak and Cane*, 1987]. The power drops off significantly towards longer periods reaching a minimum around 40 years. In the range of 40 to 400 year periods, the power ramps up, and then levels off at periods longer than 400 years. These properties of the spectrum are independent of the spectral estimation technique. The power around 4 years is ENSO, and there is a vast literature on the theory for this phenomenon [*Neelin*, 1998]. At lower frequencies, however, there is no theory. How do we assess the significance of the power at these frequencies?

The general approach in interpreting the spectra of climate records is to look for frequency bands in which the power rises above the level of a random process. Typically, the process used to define this level is an autoregressive model of order 1 (AR(1)) [*Crowley and North*, 1991]. This AR (1) model represents a linear system in which high-frequency variability is smoothed, and power at low-frequencies is emphasized. *Hasselman* [1976] described this process as integration of "weather" by the climate system. If peaks in a spectrum are above the level of the AR(1) process, they are taken to be significant at some confidence level, and can, in principle, be attributed to some deterministic process. An AR(1) process, however, is not the appropriate one against which to compare the variability in the Zebiak-Cane model. This model has a peak at a period of 4 years where most of the variance concentrated, while the AR(1) process will not allow any narrow band peaks. The next order model that is appropriate is an AR(2) process, which describes a linear, damped oscillator that is driven by white noise. Coefficients of the AR(2) are found which provide the best fit with the NINO3 index, and a 150,000 year time series is generated with white noise forcing. Comparing the spectra of the Zebiak-Cane model and the AR(2) process then is a test of whether the power at millennial timescales in the Zebiak-Cane model is significantly different than what would arise from random fluctuations of a linear process.

The spectrum of the NINO3 index and that of the AR(2) process with 95% confidence limits are shown in Figure 6. At periods longer than about 100 years, the Zebiak-Cane model has power that is distinguishable from random noise in the linear model at the 95% confidence level. We suggest that power at these frequencies is due to non-linearity in the Zebiak-Cane model.

Lorenz [1991] pointed out that power in non-linear systems can occur at unexpected low-frequencies. He writes that "...in chaotic dynamical systems in general, very-long-period fluctuations, much longer than any obvious time constants appearing in the governing laws,

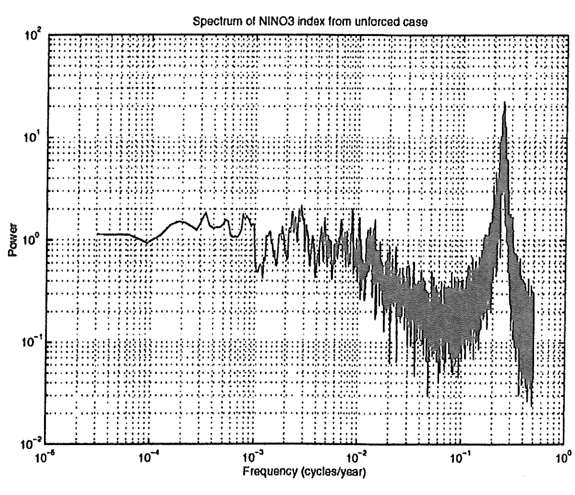

Figure 5. Multi-taper spectrum of NINO3 index from unforced Zebiak Cane model.

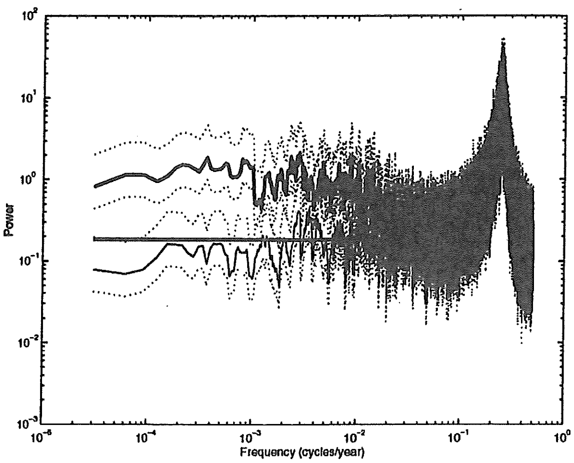

Figure 6. Multi-taper spectrum of NINO3 index from unforced Zebiak Cane model with 95% confidence limits (dotted lines around bold solid line) and multi-taper spectrum of a linear damped oscillator driven by white noise with 95% confidence limits (dotted lines around light solid line), and theoretical spectrum for an AR(2)- solid line.

are capable of developing without the help of any variable external influences." The spectrum of NINO3 can be interpreted in this context. The Zebiak-Cane model is chaotic in some regions of parameter space [*Zebiak and Cane*, 1987; *Tziperman et al.*, 1994; 1997]. The dynamics in both the atmosphere and ocean are linear and damped on a timescale of days to years [*Zebiak and Cane*, 1987]. The non-linearity arises in the coupling between the atmosphere and ocean, and thus it is the coupling that can potentially generate power at low frequencies through a non-linear cascade. It should be noted that there is no reason to expect spectral peaks from this process, yet the estimate of the spectrum of both the NINO3 index and the linear model show peaks (Figure 6). For the linear model, we know that any peak away from 4 years is purely random chance. A different realization of this processes would generate peaks at different frequencies. As for the spectrum on NINO3, if we accept that the power at low frequencies is due to this non-linear cascade, we would also conclude that, while there is significant power at low frequencies, it is over a broad band and thus the peaks there are also random.

Can we learn anything from the spatial pattern of the variability at these low frequencies? We apply a lowpass filter (> 1000 years) to the monthly fields in SST, thermocline depth and zonal and meridional wind fields, and compute the EOFs of each field. The leading EOFs (Figure 7) explain approximately 90% of the variance in each field, and the principal components are almost exactly in phase. A decreased equatorial SST gradient is accompanied by a less sloped thermocline and westerly wind anomalies on the equator. We expect these patterns to be dynamically consistent on this longer timescale because the adjustment time is on the order of months. This pattern is somewhat different from the leading ENSO EOFs of these three fields (Figure 4). In particular, there is a signal in the west Pacific that is not present on an interannual timescale, and the thermocline signal in the east Pacific is much smaller on the long timescale than the interannual timescale.

This spatial pattern is not particular to this frequency band. In fact, if we perform the same analysis on any frequency band lower than 1/50 years^{-1}, the leading EOF is the same for each of the fields. This result is consistent with the interpretation that the power at these frequencies is due to chaos. Chaotic interactions can produce self-similar behavior which gives identical patterns at different scales [*Turcotte*, 1993]. The slope of the logarithm of the power over the logarithm of the frequency is often taken as a diagnostic for the behavior of the system. It is not clear where in the spectrum to calculate the slope (there is no obvious cut-off frequency here), but the limits are -0.8 for the reddest part of

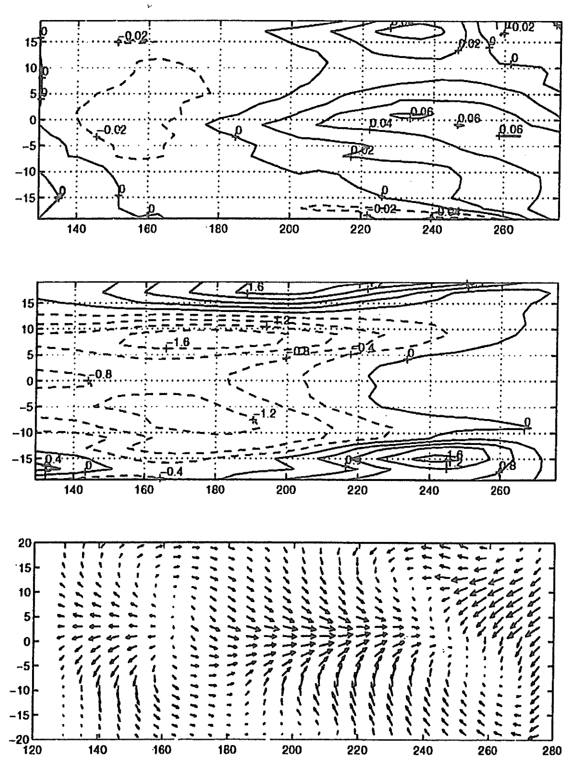

Figure 7. Leading EOF of (a) SST (K) (b) thermocline (m) and (c) wind field vectors for low pass (> 1000 yrs) unforced run.

the spectrum (1/40 - 1/400 years^{-1}), and -0.5 over the entire low frequency part of the spectrum (frequencies lower than 1/40 years^{-1}).

3. DISCUSSION

The model experiments demonstrate a mechanism by which the tropical climate can amplify Milankovitch forcing *on its own*. The mechanism is ENSO. The effect of changes in the Earth's orbital parameters in the tropics is to alter the seasonal cycle of solar radiation. ENSO is thought to be the result of interactions between the seasonal and interannual timescales [*Zebiak and Cane*, 1987; *Münnich et al.*, 1991; *Tziperman et al.*, 1994; 1997; *Jin et al.*, 1994; *Chang et al.*, 1995], and is known to be phase locked to the seasonal cycle [*Rasmusson and Carpenter*, 1982]. Thus, as the seasonal cycle in solar radiation changes, ENSO *must* be affected. The details of the response may be particular to the model, but the mechanism is not.

The results can be considered a demonstration of the idea put forth by *Palmer* [1993] that climate change in a system with strong attractors will manifest as a change in the probability that the system will reside in a particular regime without changing the character of the attractors. The primary mode of variability in the tropical climate is ENSO. The spatial structure of this mode of variability is unaltered by the solar forcing. Rather, the Milankovitch forcing changes the temporal evolution which results in a mean tropical climate change due to a change in the *statistics* of events. Again, while we can not be sure of the details, the model results presented here outline a mechanism of low frequency climate change that fits into Palmer's framework.

The model results for the unforced case raise some general issues about the interpretation of millennial scale variability in the climate record. If the real ENSO system is truly chaotic, we can expect there to be power in the climate system at low frequencies from purely tropical processes. Furthermore, this only involves near surface coupled interactions, and the deep ocean is *not* necessary for generating power at these timescales. The actual numerical results raise some particular issues. The estimate of the spectrum for both the Zebiak-Cane model and for the linear damped oscillator show peaks at low frequencies. These peaks are interpreted here to be purely random. On the basis of theory [*Turcotte*, 1993], we expect chaotic interactions to generate power in a broad band rather than at particular frequencies. We suggest that this is a more appropriate null hypothesis for testing peaks in the spectra of climate records. There may well be processes internal to the climate system that have spectral peaks at millennial timescales, but these peaks should be significant above the level of power that we expect from chaos in the climate system.

The model used in this study is highly idealized, and thus the results are subject to certain caveats. There are many processes that are not included in the model which may influence the way the coupled physics operate on Milankovitch and millenial timescales. It has been suggested that changes in the mean tropical SST and zonal SST gradient influence ENSO behavior [*Sun and Liu*, 1996, *Sun*, 1999], but many of the processes thought to be important in determining tropical SST are not included in this model. These include atmospheric convective boundary layer processes [*Sarachik*, 1978; *Betts and Ridgway*, 1989], atmospheric circulation [*Wallace*, 1992; *Fu et al.*, 1990; *Hartmann and Michelsen*, 1993; *Pierrehumbert*, 1995], ocean circulation [*Clement and Seager*, 1999], stratus clouds [*Miller*, 1997], and tropospheric water vapor content [*Larson and Hartmann*, 1999, *Seager et al.*, 1999]. On glacial-interglacial timescales, atmospheric CO_2 and high latitude climate conditions [*Broccoli and Manabe*, 1987; *Hewitt and Mitchell*, 1998] will likely affect tropical SST through both the ocean [*Bush and Philander*, 1998], and through the atmopshere [*Broccoli*, 1999]. Furthermore, changes in the strength of the Asian monsoons, either forced on a Milankovitch timescale, or unforced on a millennial timescale, are likely to have some impact on the ENSO system [i.e., *Shukla and Paolina* 1983].

The approach we have taken in this modelling study was to investigate the forced and unforced variability of this simplified coupled ocean-atmosphere system. More complete models are needed to further test the ideas presented here. The specific model results may be modified in the presence of more complete physics describing evolving climate conditions on these timescales. Nevertheless, the physical links underlying the model behavior on Milankovitch and millennial timescales are well known and unlikely to fail.

Acknowledgments. We are grateful to the two anonymous reviewers for providing useful insight. Amy Clement was supported by NASA grant NAGW-916. MAC was supported by NOAA grant NA36GP0074-02.

REFERENCES

Berger, A., Long-term variations of daily insolation and Quaternary climate changes, *J. Atmos. Sci.*, *35*, 2362–2367, 1978.

Betts, A. K., and W. Ridgway, Climatic equilibrium of the atmospheric convective boundary layer over a tropical ocean, *J. Atmos. Sci.*, *46*, 2621–2641, 1989.

Bjerknes, J., Atmospheric teleconnections from the equatorial Pacific, *Mon. Weather. Rev.*, *97*, 163–172, 1969.

Broccoli, A., Extratropical Influences on tropical paleoclimates , *AGU Monograph: Mechanisms of millenial scale global climate change*, 1998, submitted.

Broccoli, A., and S. Manabe, The influence of continental ice, atmospheric CO_2 and land albedo on the climate of the Last Glacial Maximum, *Clim. Dyn.*, *1*, 87–99, 1987.

Bush, A.B.G. and S.G.H. Philander, The role of ocean-atmosphere interactions in tropical cooling during the last glacial maximum, *Science*, *279*, 1341–1344, 1998.

Cane, M., El Niño, *Annu. Rev. Earth Planet Sci.*, *14*, 43–70, 1986.

Cane, M. A., A. C. Clement, A. Kaplan, Y. Kushnir, D. Pozdnyakov, R. Seager, S. E. Zebiak, and R. Murtugudde, Twentieth century sea surface tempature trends, *Science*, *275*, 957–906, 1997.

Chang, P., B. Wang, T. Li, and L. Ji, Interactions between the seasonal cycle and the Southern Oscillation: Frequency entrainment and chaos in a coupled ocean-atmosphere model , *Geophys. Res. Lett.*, *21*, 2817–2820, 1994.

Clement, A. C., and R. Seager, Climate and the tropical oceans, *J. Clim.*, 1999, in press.

Clement, A. C., R. Seager, M. A. Cane, and S. E. Zebiak, An ocean dynamical thermostat, *J. Climate*, *9*, 2190–2196, 1996.

Clement, A. C., R. Seager, and M. A. Cane, Suppression of el niño during the mid-holocene by changes in the earth's orbit, *Science*, 1999, submitted.

Crowley, T., and G. North, *Paleoclimatology* , Oxford Univ. Press, New York., 1991.

Dijkstra, H. A., and J. D. Neelin, Ocean-atmosphere interaction and the tropical climatology, II, Why the Pacific cold tongue is in the east, *J. Clim.*, *8*, 1343–1359, 1995.

Fu, R., A. D. DelGenio, and W. B. Rossow, Behavior of deep convective clouds in the tropical Pacific deduced from ISCCP radiances, *J. Clim.*, *3*, 1129–1152, 1990.

Goddard, L., and N. Graham, El Niño in the 1990s., *J. Geophys. Res.*, *102*, 10,423–10,436, 1997.

Hartmann, D. L., and M. Michelsen, Large scale effects on the regulation of tropical sea surface temperature, *J. Clim.*, *6*, 2049–2062, 1993.

Hasselmann, K., Stochastic climate models. Part I: Theory, *Tellus*, *XXVIII*, 473–485, 1976.

Hewitt, C. D., and J. F. B. Mitchell, A fully coupled GCM simulation of the climate of the mid-Holocene, *Geophys. Res. Lett.*, *25*, 361–364, 1998.

Intergovernmental Panel on Climate Change, *Climate change: The IPCC Scientific Assessment*, Cambridge University Press, Cambridge, England, 365pp, 1995.

Jin, F.-F., J. D. Neelin, and M. Ghil, El Niño on the Devil's staircase: annual subharmonic steps to chaos, *Science*, *264*, 70–72, 1994.

Kleeman, R., J. McCreary, and B. Klinger, A mechanism for generating ENSO decadal variability , *Science*, 1999, submitted.

Larson, K., and D. L. Hartmann, A two box equilibrium model of the tropics, *J. Clim.*, 1999, in press.

Latif, M., R. Kleeman, and C. Eckert, Greenhouse warming, decadal variability or El Niño? An attempt to understand the anomalous 1990s, *J. Clim.*, *10*, 2221–2239, 1997.

Lorenz, E., Chaos, spontaneous climatic variations and detection of the Greenhouse effect, in *Greenhouse-Gas-Induced Climate Change: A Critical Appraisal of Simulations and Observations*, edited by M. E. Schlesinger, pp. 445–453, Elsevier, Amsterdam, the Netherlands, 1991.

Mann, M., and J. Lees, Robust estimation of background noise and signal detection in climatic time series, *Clim. Change*, *33*, 409–445, 1996.

Miller, R. L., Tropical thermostats and low cloud cover, *J. Clim.*, *10*, 409–440, 1997.

Münnich, M., M. A. Cane, and S. E. Zebiak, A study of self-excited oscillations of the tropical ocean atmosphere system, II, Nonlinear cases., *J. Atmos. Sci.*, *48*, 1238–1248, 1991.

Neelin, J. D., D. Battisti, A. Hirst, F. Jin, Y. Wakata, T. Ymamgata, and S. Zebiak, ENSO theory , *J. Geophys. Res.*, *103*, 14,261–14,290, 1998.

Palmer, T. N., Extended-range atmospheric prediction and the Lorenz model, *Bull. Am. Meteorol. Soc.*, *74*, 49–65, 1993.

Pierrehumbert, R. T., Thermostats, radiator fins, and the runaway greenhouse, *J. Atmos. Sci.*, *52*, 1784–1806, 1995.

Rajagopalan, B. and U. Lall and M.A. Cane, Anomalous ENSO occurences: an alternate view , *J. Climate.*, *10*, 2351–2357, 1997.

Rasmusson, E., and T. Carpenter, Variations in tropical sea surface temperature and surface wind fields associated with the Southern Oscillation/El Niño, *Mon. Weather. Rev.*, *110*, 354–384, 1982.

Sarachik, E. S., Tropical sea surface temperature: An interactive one-dimensional model., *Dyn. Atmos. Oceans*, *2*, 455–469, 1978.

Seager, R., A. Clement, and M. Cane, Glacial cooling in the tropics: Exploring the roles of tropospheric water vapor, surface wind speed and boundary layer processes, *J. Atmos. Sci.*, 1999, submitted.

Shukla, J., and D. Paolina, The Southern Oscillation and long range forecasting of the summer monsoon rainfall over India, *Mon. Weather. Rev.*, *111*, 1830–1837, 1983.

Sun, D.-Z., Global climate change and El Niño: A theoretical framework , in *El Niño and the Southern Oscillation: Multiscale variability, global and regional impacts*, edited by E. by H.F. Diaz and V. Margraf, p. in press, Cambridge Univ. Press, New York, 1999.

Sun, D.-Z., and Z. Liu, Dynamic ocean-atmosphere coupling: a thermostat for the tropics, *Science*, *272*, 1148–1150, 1996.

Timmermann, A., J. Oberhuber, A. Bacher, M. Esch, M. Latif, and E. Roeckner, ENSO response to greenhouse warming, *Geophys. Res. Letts.*, 1998, submitted.

Trenberth, K., and T. J. Hoar, The 1990-1995 El Niño-Southern Oscillation event: Longest on record, *Geophys. Res. Lett.*, *10*, 2221–2239, 1996.

Trenberth, K., and J. W. Hurrell, Decadal atmosphere-ocean variations in the pacific, *Clim. Dyn.*, *9*, 303–319, 1994.

Turcotte, D., *Fractals and chaos in geology and geophysics*, Cambridge University Press, 1993.

Tziperman, E., M. A. Cane, and S. E. Zebiak, Irregularity and locking to the seasonal cycle in an ENSO prediction model as explained by the quasi-periodicity route to chaos, *J. Atmos. Sci.*, *52*, 293–306, 1994.

Tziperman, E., S. E. Zebiak, and M. Cane, Mechanisms of seasonal-ENSO interaction, *J. Atmos. Sci.*, *54*, 61–71, 1997.

Wallace, J. M., Effect of deep convection on the regulation of tropical sea surface temperature, *Nature*, *357*, 230–231, 1992.

Zebiak, S. E., and M. A. Cane, A model El Niño-Southern Oscillation, *Mon. Weather. Rev.*, *115*, 2262–2278, 1987.

Zebiak, S. E., and M. A. Cane, Natural climate variability in a coupled model., in *Greenhouse-Gas-Induced Climate Change: A Critical Appraisal of Simulations and Observations*, edited by M. E. Schlesinger, pp. 457–469, Elsevier, New York, 1991.

Zhang, Y., J. M. Wallace, and D. S. Battisti, ENSO-like decade-to-century scale variability: 1900-93, *J. Clim.*, *10*, 1004–1020, 1997.

M. Cane, Lamont-Doherty Earth Observatory, Palisades, New York 10964 (e-mail: mcane@ldeo.columbia.edu)

A. Clement, LODYC, Tour 26, étage 4, CASE100, 4 Place Jussieu, 75252 Paris, Cedex 05, France (e-mail:Amy.Clement@ipsl.jussieu.fr)