An Ocean Dynamical Thermostat

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ABSTRACT

The role of ocean dynamics in the regulation of tropical sea surface temperatures (SSTs) is investigated using the Zebiak–Cane coupled ocean–atmosphere model. The model is forced with a uniform heating, or cooling, varying between ±40 W m^{-2} into the ocean surface. A new climatological SST pattern is established for which the area-averaged temperature change is smaller in magnitude than the imposed forcing. The forcing is balanced almost equally by a change in the heat flux out of the ocean and by vertical advection of heat in the ocean through anomalous equatorial ocean upwelling. The generation of anomalous upwelling is identified here as a possible mechanism capable of regulating tropical SSTs. This ocean dynamical thermostat mechanism has a seasonally varying efficiency that causes amplification (weakening) of the seasonal cycle for the heating (cooling). The interannual variability also changes under the imposed forcing. These results suggest that the role of ocean dynamics should be included in any discussion of the regulation of the tropical climate.

1. Introduction

The presence of water vapor in the tropical atmosphere has a potentially destabilizing effect on the tropical climate. As sea surface temperatures (SSTs) increase, the ocean attempts to cool through evaporation. This, however, increases the water vapor in the atmosphere, which traps longwave radiation. At high enough SSTs the radiative loss to space from clear skies can actually decrease with increasing SST (Hallberg and Inamdar 1993), a situation referred to as the supergreenhouse effect. Ramanathan and Collins (1991, hereafter RC) postulated that cirrus clouds act as a brake on this positive feedback. They suggested that cirrus cloud cover associated with deep convection increases locally over the warm pool as the SSTs increase, having the net effect of blocking out incoming solar radiation. They identified this as a thermostat mechanism that would limit the warm pool SSTs to 305 K. Ramanathan and Collins claimed that this hypothesized thermostat explains the observed negative skewness of the tropical SST distribution. Wallace (1992) pointed out that large-scale atmospheric dynamics requires tropical tropospheric temperatures to be fairly uniform and that efficient ocean–atmosphere heat exchange in regions of deep convection would lead to the observed negative skewness in tropical SSTs in the absence of the radiative effects of cirrus clouds. The thermostat has also

been criticized by Fu et al. (1992) on the basis that the location of deep convection, and the surface energy budget, are related not so much to local SSTs as to the large-scale atmospheric circulation. Sarachik (1978) and Betts and Ridgway (1989) also show that the tropical climatology is determined nonlocally.

Newell (1979) and Hartmann and Michelsen (1993) proposed a different thermostat that operates through evaporative feedback. Newell (1979) argued that above 30°–31°C, evaporative cooling would exceed the heat input from radiation and that tropical SSTs are thus limited. He did not, however, consider the positive feedback of enhanced greenhouse trapping that accompanies increasing SST due to increased atmospheric water vapor content. Hartmann and Michelsen (1993) argue that the stabilization of SSTs due to evaporative cooling can overcome this positive feedback. Using a simple heuristic model, they show that the large-scale atmospheric circulation responds to SST gradients, and the vigor of this circulation can supply dry air to the boundary layer, allowing for evaporative cooling of the ocean surface. It follows that the mean tropical SSTs are coupled to the SST gradients in the Tropics. Furthermore, they found that averaged over the area of the large-scale circulation that contains deep convection, there is little sensitivity of cloud albedo to SST changes, thus disputing RC's cloud thermostat.

Pierrehumbert (1995) has proposed perhaps the most unified theory to date. He begins with the as yet unexplained observation that clouds have no net effect on the top of the atmosphere radiation and follows this idea to its logical conclusion to demonstrate that deep convective clouds can have no net effect in stabilizing the tropical climate. His theory instead relies on the

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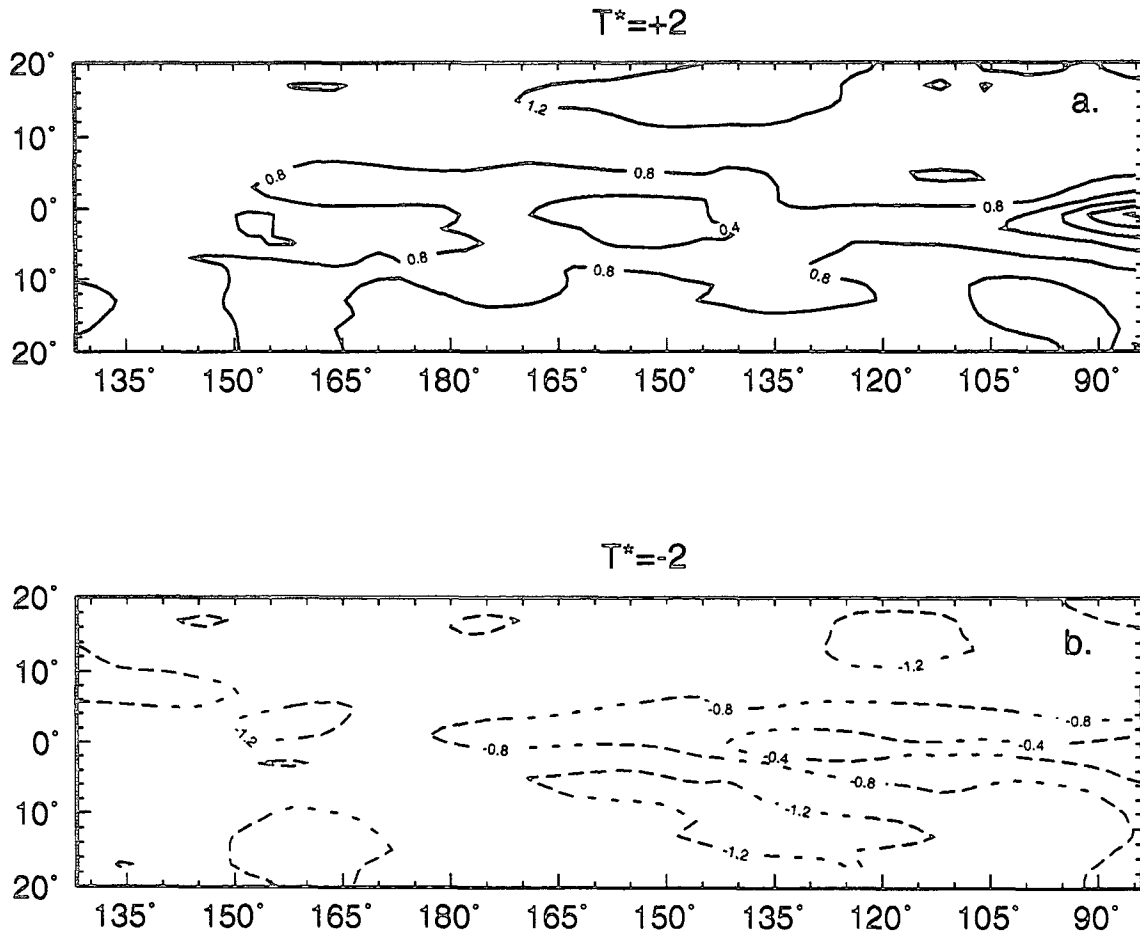## April SST anomaly

T*=+2

T*=-2

FIG. 1. (a) SST anomalies four model months after the start of the run for the $T^* = +2$ case. (b) Same as (a) but for the $T^* = -2$ case. Solid contours are positive anomalies and dashed are negative.

presence of "radiator fins" through which the atmosphere can rid itself of excess energy. Because of greenhouse trapping, the atmosphere, unable to rid itself of this energy in the regions of deep convection (the "furnace"), instead exports energy to the drier, nonconvecting regions of the subtropics (the "radiator fins") where it can be effectively radiated to space. Pierrehumbert identifies the relative sizes of the cold pool–warm pool configuration, as well as the dryness of the radiator fins, as the main factors in determining the tropical climate.

One aspect these highly divergent views on the regulation of tropical SST have in common is the lack of any interactive dynamical transports of heat in the ocean. In discussing CO_2-induced climate change, RC speculate that their thermostat would leave warm pool SSTs relatively unchanged, whereas SSTs in regions outside the regions of deep convection would warm. Hartmann and

Michelsen (1993) specify the changes in the SST gradients in their model. Pierrehumbert (1995) specifies the ocean heat transports and the relative sizes of the warm pool (A_1) and the cold pool (A_2), while speculating that there is "an undiscovered stabilization principle at work which involves adjustment of A_2/A_1." Hence all of these thermostat mechanisms involve changes in tropical SST gradients and, by implication, the low-level winds. However, they ignore the fact that changes in the atmospheric circulation will alter the ocean circulation that will further feedback to the SSTs. Instead, they take the ocean to be dynamically inactive, entering only as a thermodynamic layer.

In this study, we illustrate a possible role for ocean dynamics in regulating tropical SSTs by including only highly idealized atmospheric thermodynamics. Using a coupled ocean–atmosphere model, we perform the simplest experiment possible and find that ocean dynamics

not only alter the mean tropical SST and climatology but also affect the amplitude of the seasonal cycle and the interannual variability. This suggests that ocean dynamics cannot be ignored in the discussion of the stability of the tropical climate and can be a complicating but essential element in climate change studies.

2. Experiment with ZC model

We use the Zebiak–Cane model (Zebiak and Cane 1987, hereafter ZC), which is a coupled ocean–atmosphere model that solves for perturbations about the climatological state. The model consists of an atmosphere governed by linear shallow-water equations on an equatorial beta plane and a linear reduced-gravity ocean model (for more details see ZC). The model domain extends from 29°N to 29°S, 124°E to 80°W. The temperature anomaly T in the ocean model mixed layer is determined from the following equation (written in flux form):

$$\frac{\partial T}{\partial t} = -\nabla \cdot (\tilde{\mathbf{u}}\tilde{T} - \bar{\mathbf{u}}\bar{T})$$

$$+ \left(\frac{\tilde{w}_e \tilde{T}_e}{H} - \frac{\bar{w}_e \bar{T}_e}{H} \right) - \alpha T + \alpha T^*, \quad (1)$$

where T_e is a nonlinear function of the subsurface temperature, the surface temperature, and w_e [cf. ZC Eq. (1)]. This function accounts for the fact that the surface temperature is only affected by vertical advection in the presence of upwelling. The tilde indicates total (mean plus anomaly) quantities; the overbar denotes the mean quantities; quantities without this subscript are anomalies; the subscript e indicates values at the base of the mixed layer; H is the mixed-layer depth (50 m); α^{-1} is a timescale. The term involving T^* represents a uniform forcing implemented as a constant heat flux into the ocean that varied among runs between approximately ±40 W m^{-2} (α is set to 125 days^{-1} and T^* was varied between ±2°C). This experiment is intended to simulate how the coupled tropical system would respond to a simple forcing. In the absence of ocean dynamics, the model would generate an SST anomaly $T = T^*$.

Figures 1a and 1b show the surface temperature anomaly in April, four model months after the start of the runs. The temperature change in the eastern equatorial region (approximately 180° to the eastern boundary and between 5°S and 5°N) is less than that of the surrounding regions. Seager et al. (1988) pointed out that the ocean's response to a uniform heating is not spatially uniform. In the western equatorial Pacific and the off-equatorial regions, the SST is primarily determined through a one-dimensional balance between heat storage and the surface flux. Here the SST must increase, or decrease, such that the surface heat flux anomaly balances the imposed forcing. In the eastern

equatorial region, the imposed forcing can instead be partially balanced by anomalous horizontal and vertical advection, and the SST will change less.

In the positive T^* case, for example, this means that the east–west temperature gradient is increased. This will strengthen the equatorial easterlies which, in turn, will increase the upwelling in the central equatorial Pacific and cause the thermocline to shoal in the east. Both processes will further cool the SSTs in the eastern part of the basin, leading to a coupled interaction that establishes a new climatology. The resulting annual mean SST anomalies are shown in Figs. 2a and 2b for the warming and cooling cases. Rather remarkably, the coupled interaction actually causes the temperature anomaly in the NINO3 region (5°S–5°N, 90°–150°W) to be of the opposite sign to that of the forcing. As a consequence, the basin mean temperature changes less than would be the case if the ocean had equilibrated through the surface heat flux alone. Figure 3 shows the terms in Eq. (1) averaged over the area of the entire basin. The forcing, αT^*, is almost equally balanced by the change in heat flux and the vertical advection of temperature.

What we have identified in this experiment is a tropical ocean dynamical thermostat. An imposed heating induces greater equatorial upwelling that cools the equatorial SST. Meridional advection spreads the upwelled water off the equator leading to a basin average temperature change that is less than expected (Fig. 4).

Because this is an anomaly model, however, it is not energetically closed. The mean vertical thermal structure is specified such that the basic state provides an infinite source or sink of energy. In reality, the thermocline temperature would change, with a substantial lag, as the SST changes in the subduction regions north and south of the equator. The sources of the upwelled water are only beginning to be understood (Liu et al. 1994; Lu and McCreary 1995), and determination of how the subsurface temperature is modified would require a more complete ocean–atmosphere model. Adjustment of the thermocline would presumably damp the effectiveness of the thermostat but not alter its basic character. The changing subsurface temperature would also introduce another, longer, timescale to the coupled dynamics. This issue is currently being investigated using a high-resolution ocean model.

In addition to the changes in the annual mean SST field, the ocean dynamics have an unexpected effect on how the amplitude of the seasonal cycle and interannual variability responded to a uniform forcing. Equatorial upwelling is at its weakest in the Northern Hemisphere spring when the SST is at a maximum and the ITCZ is closest to the equator, and it is at its strongest in Northern Hemisphere fall when the SST is a minimum and the ITCZ is well north of the equator. Thus, the efficiency of the ocean dynamical thermostat has a seasonal cycle. In Northern Hemisphere spring it can only weakly oppose the forcing. In contrast, during the

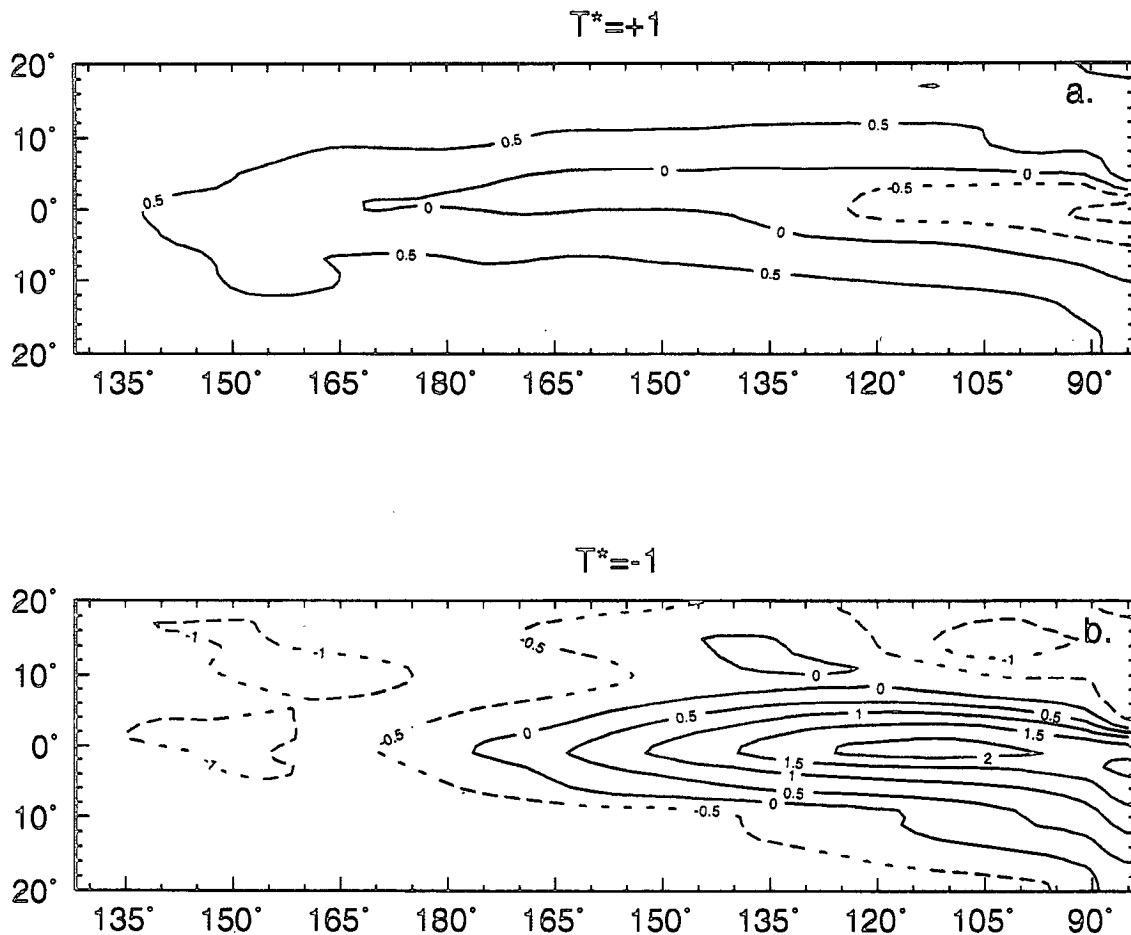## Annual mean SST anomaly

T*=+1

T*=-1

FIG. 2. (a) Annual mean SST anomalies for the $T^* = +1$ case. (b) $T^* = -1$. Solid contours are positive anomalies and dashed are negative.

fall, the strong upwelling efficiently opposes the forcing. Thus, the SST response is smaller in magnitude in the spring than it is in the fall. Because the sign of the response in this region is opposite to that of the forcing, this constitutes an enhancement of the seasonal cycle about the new mean NINO3 anomaly for the warming cases and a weaking for the cooling cases (Fig. 5). Generation of anomalous upwelling will intensify this response. This happens for even a modest warming or cooling ($T^* = \pm 0.5$). The amplitude of the background seasonal cycle in the NINO3 region can change by almost one-third for the $T^* = \pm 2$ case.

The interannual variability changes dramatically over the range of T^* (Fig. 6). For the warming, the ENSO variability is almost completely damped out, while for the cooling the events become more regular. These changes are related to the shape of the vertical temperature profile. The subsurface temperature in the

model is parameterized as a hyperbolic tangent function of the thermocline depth anomaly (Zebiak and Cane 1987). When the thermocline depth anomaly is close to zero, the slope of the curve is large (i.e., a small change in thermocline depth will result in a large change in subsurface temperature). This causes the chaotic activity in the standard ($T^* = 0$) case. In the $T^* > 0$ case, however, the thermocline is shallower in the east relative to the mean, and the coupling between subsurface temperature and thermocline motions is weaker. The result is damping of the interannual variability, so that almost all of the power is in the annual cycle. Chang et al. (1995) have examined the interaction between ENSO and the seasonal cycle and found similar results. Varying the strength of the seasonal cycle in surface heat flux, they find that ENSO variability undergoes period doubling bifurcations that can lead to chaos. When the seasonal cycle is amplified by a factor
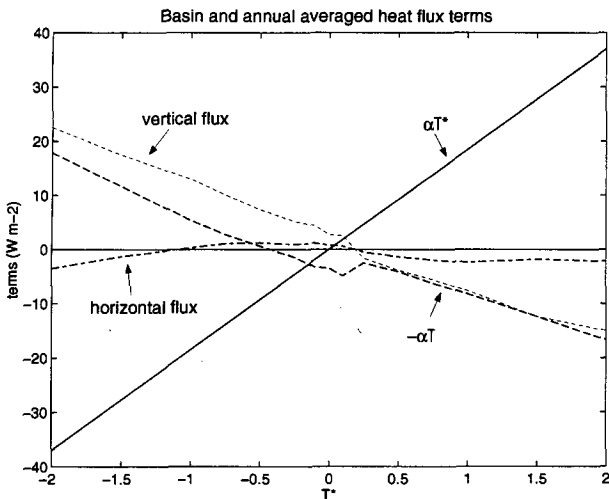
FIG. 3. Climatology (long-term mean) of the terms in Eq. (1) averaged over the entire basin (W m^{-2}) as a function of T^*. Solid: αT^*; dashed: αT; dotted: change in vertical flux (upwelling); dot–dashed: change in horizontal flux.
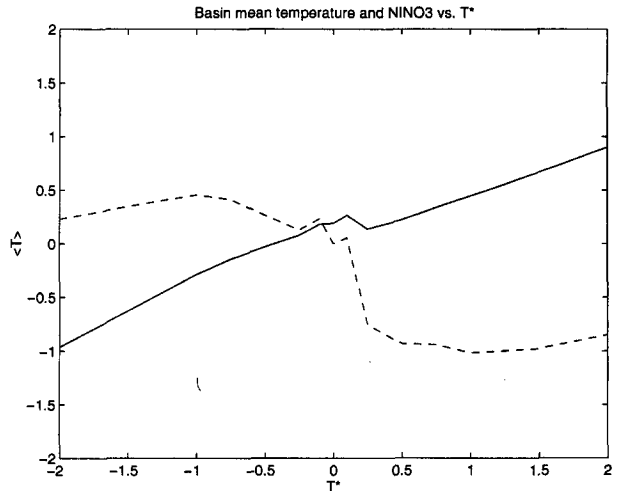

FIG. 4. Basin and annual mean temperature anomaly (solid) and NINO3 temperature anomaly (dashed) relative to the standard ($T^* = 0$) run as a function of T^*.

of 1.27, the variability becomes locked into the annual cycle. The extent to which the changes in the interannual variability seen in this experiment and that of Chang et al. (1995) would persist in the real world depends on how the vertical thermal structure adjusts through advective and diffusive processes.

3. Discussion and conclusions

We have taken an admittedly simple model and subjected it to an external heat flux forcing that is the most idealized possible. The model takes account of atmospheric thermodynamics only to the extent that they directly affect the low-level winds. The requirement for atmospheric energy balance is ignored. This experimental arrangement is the opposite extreme to that of the recent thermostat theories. Such theories gave the ocean no role and found a wide variety of behavior; we allow the atmosphere little freedom and find powerful control on the climate system from ocean dynamics. The ocean dynamics have a thermostat effect such that the basin mean SST response is about one-half that of the imposed forcing. This mechanism involves a different configuration of SSTs where the east–west and north–south gradients differ from the modern climatology. As the theories on atmospheric thermostats show, both of these would have a significant effect on the large-scale atmospheric circulation.

It is interesting to speculate how the ocean thermostat would be modified in the presence of more complete atmospheric thermodynamics. Consider the warming case ($T^* > 0$), where the ocean dynamics cause the zonal equatorial SST gradient to increase while the meridional gradient decreases (Fig. 2). This implies a change in the relative strengths of the Hadley

and Walker circulations. If this leads to a decrease of the area of deep convection, Pierrehumbert's model predicts that the atmosphere would radiate more effectively since the radiator fin area increases, further stabilizing the tropical climate. The opposite is the case if the convective area increases in response to the changes in large-scale atmospheric circulation and would lead to a runaway warming. Pierrehumbert's theory, however, is at odds with the observational analysis of Chou (1994). Chou found that during the warm ENSO phase when the radiator fin area is smaller, the earth loses
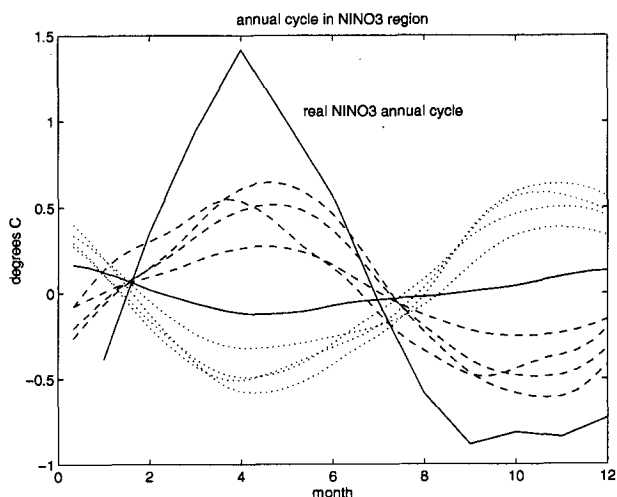

FIG. 5. Seasonal cycle of surface temperature anomaly for the NINO3 region for values of $T^* = \pm 0.5, 1, 1.5, 2$. Dashed lines are for $T^* > 0$; dotted lines are for $T^* < 0$; solid line is $T^* = 0$. The observed annual cycle of SST for the NINO3 region is also shown (the long-term mean is first subtracted).
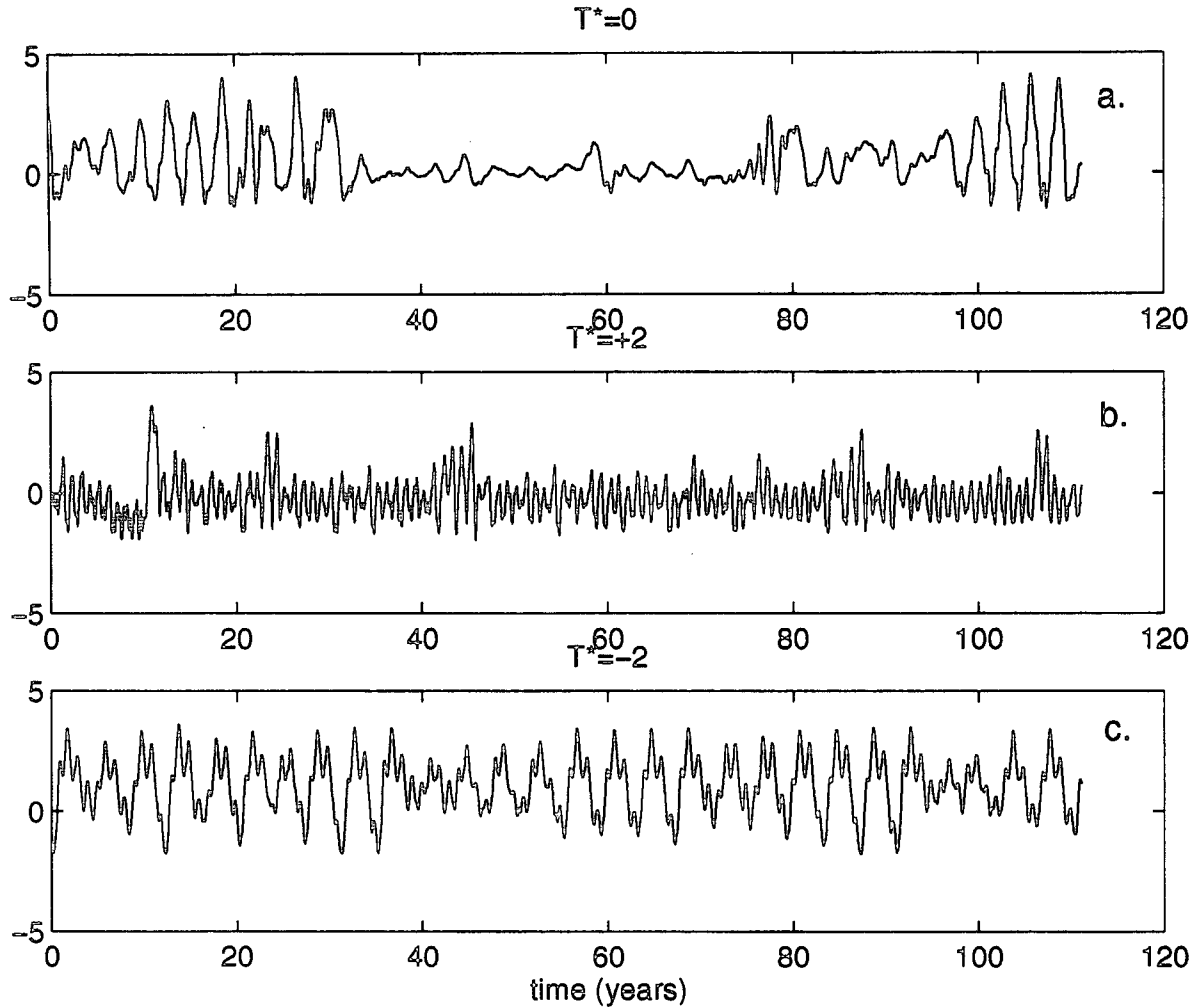
FIG. 6. Representative segments of time series of the NINO3 index taken from a 1000-yr run for
(a) $T^* = 0$, (b) $T^* = +2$, and (c) $T^* = -2$.

more radiant energy to space, whereas Pierrehumbert's model would predict that with a smaller radiator fin, the earth loses less energy to space. Chou demonstrated that this is due in part to changes in SST and in part to the clear-sky radiation responding to changes in the upper-tropospheric water vapor content. Pierrehumbert ignores this latter effect by fixing the radiator fin emissivity. Another highly idealized aspect of the thermodynamics in the ZC model is that the surface heat flux is related only to the surface temperature anomaly and a uniform damping timescale. In reality, the sensitivity of surface heat flux to SST is spatially variable, and Seager et al. (1995) show it to be larger in the subtropics than it is in the Tropics due to higher wind speeds there. Because of this, the ZC model may overestimate the subtropical SST changes. Hartmann and Michelsen (1993) have demonstrated the coupling be-

tween SST gradients, atmospheric circulation, and mean tropical SST. What they find when the zonal SST gradient is increased (i.e., $T^* > 0$ cases) is a more vigorous atmospheric circulation that causes wind-induced latent heating anomalies and reduces the mean tropical SST. This would enhance the thermostat effect. Considering these different components of the atmospheric thermodynamics in this noninteractive manner, it appears to be more likely that the coupling with the atmosphere will enhance the ocean thermostat. However, we are keenly aware that such speculation overlooks many possible additional feedbacks that could control the response. This issue should be addressed with a more complete account of the atmosphere. What clearly must be considered as well is that the ocean SSTs are not determined solely by the atmosphere. As we have shown, the ocean has its own way of managing

anomalous heating or cooling, which provides a compelling argument for the role of the ocean in the regulation of the tropical climate.

In discussing the problems in coupled models introduced by artificial flux corrections, Dijkstra and Neelin (1995) have found that a uniform flux error can, through coupled dynamics, give an error field that has spatial structure. The mechanisms are the same as those discussed here. In contrast, a study by Knutson and Manabe (1995) using a coupled GCM with a coarse-resolution ocean shows a more or less uniform change in tropical SST for quadrupled CO_2, but also a small reduction in the equatorial SST gradient. This is opposite to the results of the warming experiments performed here where the SST gradient increased. Knutson and Manabe attribute this reduction to an enhancement of the evaporative damping of the zonal SST gradient with ocean dynamics having no net effect. However, the reduced gradient only occurs after 100 years of integration (T. Knutson, personal communication, 1996). Apparently, the mechanism presented here, which would cause an initial increase in the SST gradient, is not operative in their model. We suspect that the discrepancy between these results lies in differences in the coupling of SST to the equatorial thermocline and the subsurface temperature. This issue is currently being addressed with a high-resolution ocean GCM. Since the response of ocean dynamics to a uniform forcing affects not only the basin mean temperature but also the pattern of SST change, the seasonal cycle, and the interannual variability, this issue must be resolved before one can have confidence in the ability of coupled GCMs with low-resolution ocean models to assess real climate variability.

The forcing we impose in these experiments is highly idealized, and in reality the forcing associated with any climate change scenario—be it glacial changes or greenhouse warming—will be more complicated. We view these experiments as representing the simplest possible perturbation of the coupled system. Because of ocean dynamics, even the simplest forcing involves changes in the pattern of annual mean SST, the seasonal cycle, and interannual variability, all central issues to both the tropical and global climate. The response to a more realistic forcing would undoubtedly be more complex. These simple experiments make it clear that any theory of tropical stability or climate change is incomplete if it ignores the role of equatorial ocean dynamics.

REFERENCES

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

Chang, P., B. Wang, T. Li, and L. Ji, 1995: Interactions between the seasonal cycle and the Southern oscillation—Frequency entrainment and chaos in a coupled ocean–atmosphere model. *Geophys. Res. Lett.,* in press.

Chou, M., 1994: Coolness in the tropical Pacific during an El Niño episode. *J. Climate,* **7,** 1684–1692.

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

Fu, R., A. D. Del Genio, W. B. Rossow, and W. T. Liu, 1992: Cirrus-cloud thermostat for tropical sea surface temperatures tested using satellite data. *Nature,* **358,** 394–397.

Hallberg, R., and A. K. Inamdar, 1993: Observations of seasonal variations in atmospheric greenhouse trapping and its enhancement at high sea surface temperatures. *J. Climate,* **6,** 920–931.

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

Liu, Z., S. G. H. Philander, and R. C. Pacanowski, 1994: A GCM study of tropical-subtropical upper-ocean water exchange. *J. Phys. Oceanogr.,* **24,** 2606–2623.

Lu, P., and J. P. McCreary, 1995: Influence of the ITCZ on the flow of thermocline water from the subtropical to the equatorial ocean. *J. Phys. Oceanogr.,* **25,** 3076–3088.

Knutson, T. R., and S. Manabe, 1995: Time–mean response over the tropical Pacific to increased CO2 in a coupled ocean–atmosphere model. *J. Climate,* **8,** 2181–2199.

Newell, R. E., 1979: Climate and the ocean. *Amer. Sci.,* **67,** 405–416.

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

Ramanthan, V., and W. Collins, 1991: Thermodynamic regulation of ocean warming by cirrus clouds deduced from observations of the 1987 El Niño. *Nature,* **351,** 27–32.

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

Seager, R., S. Zebiak, and M. A. Cane, 1988: A model of the tropical Pacific sea surface temperature climatology. *J. Geophys. Res.,* **93,** 1265–1280.

——, Y. Kushnir, and M. A. Cane, 1995: On heat flux boundary conditions for ocean models. *J. Phys. Oceanogr.,* **25,** 3219–3230.

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

Zebiak, S. E., and M. A. Cane, 1987: A model El Niño–Southern oscillation. *Mon. Wea. Rev.,* **115,** 2262–2278.