NOTES AND CORRESPONDENCE

On Heat Flux Boundary Conditions for Ocean Models*

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ABSTRACT

Recent modeling studies of thermohaline variability have imposed rapid damping of modeled sea surface temperature (SST) anomalies equivalent to assuming the atmosphere has an infinite heat capacity. Such surface heat flux parameterizations effectively exclude the possibility of SST playing an active role in the thermohaline circulation. The authors present results of simple thermodynamic modeling of the lower atmosphere that suggest the sensitivity of the surface heat fluxes to variations in SST is much smaller than often assumed. It is found that the flux response is strongly dependent on the scale of the SST anomaly. For the very largest scales the fluxes increase by only a few watts per square meter per kelvin change of SST. For the scales typical of observed anomalies the nonlocality of the response enhances the sensitivity, which may reach up to ~15 W m^{-2} K^{-1}. This extreme is still less than half of the values typically assumed in ocean models. The small sensitivity arises from the adjustment of the lower atmosphere to the underlying ocean in accord with its relatively much smaller ability to store heat and moisture. The increase in fluxes with SST is dominated by the latent heat flux but offset significantly by reduced net longwave radiative cooling of the surface.

1. Introduction

Variability of the thermohaline circulation has received increasing attention in recent years. Numerous workers have found that oscillations of the thermohaline circulation can be produced in ocean models in the absence of variable wind forcing. Multiple equilibrium states of the circulation have also been found under the same external forcing and different initial conditions. These results have led to speculation that ocean models can reproduce the deduced observed variability with some degree of realism.

The surface flux boundary conditions most commonly used in these studies are referred to as "mixed boundary conditions." This combines an imposed freshwater flux with a restoration of the models' surface temperature to some assumed value that varies in space but not in time (e.g., Marotzke and Willebrand 1991; Weaver and Sarachik 1991a,b; Winton and Sarachik 1993). These models variously produced oscillations and multiple equilibria of the thermohaline circulation. The oscillations were closely linked to variations in the modeled salinity fields, which influenced the stability of the water column and, hence, the formation of deep water. Weaver et al. (1993) have attempted to unify the differing results and have shown that differing behavior can be obtained from the same model subject to slightly different salinity forcing. Further, Tziperman et al. (1994) have illustrated that the stability of the thermohaline circulation depends on the assumptions made in setting the boundary conditions on salinity.

All of these studies restore the model sea surface temperature (SST) to fixed temperatures on a timescale of a few tens of days. This corresponds to a variation of the surface heat flux with the SST on the order of several tens of W m^{-2} K^{-1}. These values correspond to those suggested by Oberhuber (1988) and Haney (1971). The combined effect of a strong damping of SST anomalies and the fixed restoring temperature negates the possibility that SST can play an independent role in the variability.

In this note we apply a model of the advective atmospheric mixed layer to address the question of what is the correct relationship between SST variability and flux variability. The model uses a set of quasi-equilibrium assumptions to derive surface fluxes on the basis of winds, SST, and cloud cover alone. It is designed for coupling to ocean models to allow the model to determine its own SST taking full account of the feedbacks that exist between the fluxes and the SST. In the next section we provide a brief review of thermal boundary conditions used by ocean models; in section

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3 we describe the atmospheric mixed layer model; results are presented in section 4; a discussion of the implications is in section 5; and conclusions are given in section 6.

2. Review of heat flux boundary conditions used in ocean models

The most commonly used boundary condition is to restore the SST to an equilibrium value as

$$Q = \kappa(T_0 - T_a). \tag{1}$$

Here Q is the net surface heat flux (including all components and defined as positive upwards), κ is a coupling coefficient, T_0 is the SST, and T_a is the restoring temperature. Marotzke and Willebrand (1991) used this formulation and assumed a T_a that closely tracked the observed zonal mean distribution of SST. Weaver and Sarachik (1991a) used the same formulation with T_a derived from the observed zonal SST data. The coupling coefficient has units of watts per square meter per kelvin. A timescale can be derived as $\tau = \rho c_p h/\kappa$, where ρ is the density, c_p is the specific heat of seawater, and h is a typical mixed layer depth. Weaver and Sarachik (1991a) used a timescale of 25 days. Marotzke and Willebrand (1991) used 30 days. For a mixed layer 50 m deep these correspond to values of κ of 97 W m^{-2} K^{-1} and 81 W m^{-2} K^{-1}, respectively.

A heat flux formulation of this type presents three obvious problems. The first involves the distribution of mean heat flux. The observed annual-mean net heat flux (including all terms) is observed to be into the ocean in low latitudes and from the ocean to the atmosphere in higher latitudes (e.g., Esbensen and Kushnir 1981). According to (1) large heat fluxes can only occur if the SST differs significantly from the restoring temperature. Most workers nonetheless take T_a to be something close to the observed SST because this forces the model SST to be close to that observed. Of course, over a sufficient timescale, the integral over space of the heat flux must be zero, requiring that the meridional gradient of SST be less than that of T_a. But, to the extent that the model is successful in producing the observed zonal-mean SST (i.e., T_0 approaches T_a), the heat flux goes to zero, which is incorrect. Moreover, locally, a zero heat flux implies no heat transport by the ocean circulation. Hence it is pointless to look at ocean heat transport in any model that restores the SST to observed values. With a heat flux given by (1), and T_a close to the observed SST, it is impossible to get the correct combination of SST, heat fluxes, and oceanic heat transport.

Second, this boundary condition also assumes the atmosphere has an infinite heat capacity because, whatever the SST is, the restoring temperature does not change. A wealth of evidence is available to the contrary, not the least of which the close correlation between the observed SST and air temperature (e.g., Esbensen and Kushnir 1981; Weare et al. 1980). An infinite heat capacity atmosphere acts as an infinitely large local source or sink of energy for the ocean. It ignores the fact that the atmosphere will adjust quickly to the underlying SST. This point has been made by Schopf (1983) and Rahmstorf and Willebrand (1995, RW hereafter). The same criticism holds equally for moisture: it ignores the adjustment of the atmospheric humidity field to evaporation at the ocean surface.

The third problem involves the magnitude of the coupling coefficient κ. Use of large values of κ is sometimes justified by reference to the values for κ presented in the heat flux atlas of Oberhuber (1988). However, though not explicitly stated, it appears that Oberhuber also assumed that the atmosphere did not change in response to SST changes. He calculated values of κ by differentiating the bulk formulas for heat flux with respect to SST and assuming the air temperature and air humidity remained constant. This gives values upward of 40 W m^{-2} K^{-1} that are dominated by the increase of saturation humidity with SST. Assuming a fixed relative humidity RH, instead would reduce this estimate by a factor of $(1 - \mathrm{RH})$, that is, to about 10 W m^{-2} K^{-1} for a reasonable RH of 0.75.

Rahmstorf and Willebrand recognized some of these problems and derived a different heat flux boundary condition on the basis of the heat budget of the atmosphere. This allows the atmospheric temperature to adjust to the underlying SST. The timescale for adjustment of the SST by heat fluxes is then set by the ability of the atmosphere to adjust to perturbations in the surface energy flux through radiative loss to space. They derive a value of κ of only 2 or 3 W m^{-2} K^{-1} corresponding to a timescale of years. Their heat flux formula still includes a term with the form of (1), although the interpretation of T_a is somewhat different. They also include a diffusive term that parameterizes atmospheric dynamics.

Rahmstorf and Willebrand use a zonally uniform T_a that varies from a few kelvin warmer than the observed SST in the Tropics to a few kelvin colder at high latitudes. Combined with a small κ, the local term in their heat flux formulation provides rather small net heat fluxes. However, RW formulated their heat flux in such a way that the nonlocal, diffusive, term makes up the difference and provides net fluxes on the order of magnitude observed (S. Rahmstorff 1995, personal communication). They defined T_a as the temperature the ocean would adjust to if it did not transport heat. The diffusive term represents the changes in the net surface heat flux that arise from differences in the atmospheric circulation between that state and the realistic state in which the ocean does transport heat. The Rahmstorf and Willebrand method appears to be more realistic than the simple restoring condition of (1). However, it relies on an idealized atmospheric energy budget and ignores zonal asymmetries.

An alternative approach to computation of heat fluxes is to use some assumption about turbulent processes in the atmospheric boundary layer to derive the near-surface air humidity and temperature in terms of the SST and winds. This was done for the tropical Pacific by Seager et al. (1988). They assumed that local, one-dimensional equilibrium prevailed such that the air humidity adjusted to a fixed proportion of the saturation humidity evaluated at the SST. This proved successful in a simulation of the tropical Pacific SST. The method was extended by Blumenthal and Cane (1989) and Seager and Blumenthal (1994). Nonetheless, this scheme was invalid in regions of strong advection where one-dimensional equilibrium did not hold. This made it inappropriate for simulating the large fluxes of latent and sensible heat off the wintertime Northern Hemisphere continents.

To overcome this problem Seager et al. (1995, SBK hereafter) have developed a model of an advective atmospheric mixed layer that explicitly calculates the near-surface air temperature and humidity needed to calculate the surface fluxes. The winds must be externally prescribed. Thus, here we address only the thermodynamic response of the lower atmosphere and leave aside consideration of changes in the total heat flux due to atmospheric circulation changes that may modify the wind speed or solar radiation. Nonetheless, this model provides a tool with which to examine the feedbacks between fluxes and SST. This is consistent with what is needed for a boundary condition by ocean models, which are typically run with specified winds, solar radiation, and cloudiness.

3. The advective atmospheric mixed layer model

The model, described in detail in SBK, seeks to represent either a dry convective layer or the subcloud layer that underlies marine clouds. Within this layer it determines the virtual potential temperature and specific humidity by balancing advection, diffusion, the fluxes at the surface and the mixed layer top, and, for temperature, radiative cooling. The model assumes a steady state because of the rapid timescale, less than a day, on which the mixed layer adjusts to changes in surface fluxes.

The surface fluxes are computed using the usual bulk formula. The closure for the flux of virtual potential temperature at the mixed layer top has been justified on the basis of data analysis (Nicholls and LeMone 1980), modeling (Betts 1976), and theory (Tennekes 1973). It sets the downward flux at the mixed layer top to be a fixed proportion of the surface flux and has been used extensively in models of marine boundary layers (e.g., Bretherton 1993; Betts and Ridgway 1989; Albrecht et al. 1979). The radiative cooling is assumed to be a constant 2 K day^{-1}.

The closure for the moisture flux is more empirical and relates the turbulent flux at the mixed layer top to the generation of turbulence at the surface by friction. The moisture above the mixed layer is taken to be a fixed proportion of the mixed layer humidity. This closure preserves the observed correlation between surface moisture flux and wind speed.

With these assumptions the model equations are (see SBK for a complete derivation)

$$P\mathbf{u}\cdot\nabla\theta_v = (1+\beta_V)C_0\omega_0(\theta_{V0}-\theta_V) + P\nu\nabla^2\theta_V + PR', \tag{2}$$

$$P\mathbf{u}\cdot\nabla q = C_0\omega_0 q_0 - C_0\omega_0(1+\mu)q + P\nu\nabla^2 q, \tag{3}$$

$$\theta = \theta_V/(1+.61q), \tag{4}$$

where P is the fixed mixed layer pressure thickness (a typical value for the subcloud layer of 60 mb is assumed), θ_V is the virtual potential temperature and θ_{V0} is its surface value, q is the specific humidity, and q_0 is the saturation specific humidity at the surface temperature; θ is the potential temperature, ν is a diffusion coefficient, R' is $(1+.61q)$ times the radiative cooling, C_0 is the surface exchange coefficient, and ω_0 is a surface velocity scale; β_V is the closure parameter that determines the virtual potential temperature flux at the mixed layer top (see Betts 1976) and μ is a parameter related to the closure on the moisture flux at the mixed layer top. Here μ is set so that, in local equilibrium [$q = q_0/(1+\mu)$], the modeled relative humidity is close to the observed value of 80%.

Observed values of virtual potential temperature and humidity [from the European Center for Medium-Range Weather Forecasts (ECMWF) analyses averaged over the period 1985–1992] are specified around the continental margins and the advection–diffusion equations, (2) and (3), are solved to obtain steady-state solutions subject to these boundary conditions. The advecting winds are 1000-mb winds analysed by ECMWF. Once virtual potential temperature and humidity are known, the temperature can be derived from (4).

SBK present global simulations of the sensible and latent heat fluxes obtained by this model. The results are in general satisfactory and reproduce all the important observed features, though the enhanced fluxes over the Kuroshio and Gulf Stream in winter are weaker than observed. They suggest that the remaining quantitative problems can be traced to the assumption of a constant depth mixed layer.

4. Sensitivity of the surface heat flux to SST perturbations

a. *Sensitivity to a globally uniform change in SST*

The first case that we consider is for a globally uniform change in SST. The heat fluxes are calculated for a uniform increase of 1 K in the SST relative to climatological monthly mean values, and then for a uniform decrease of 1 K. Dividing the change in heat flux

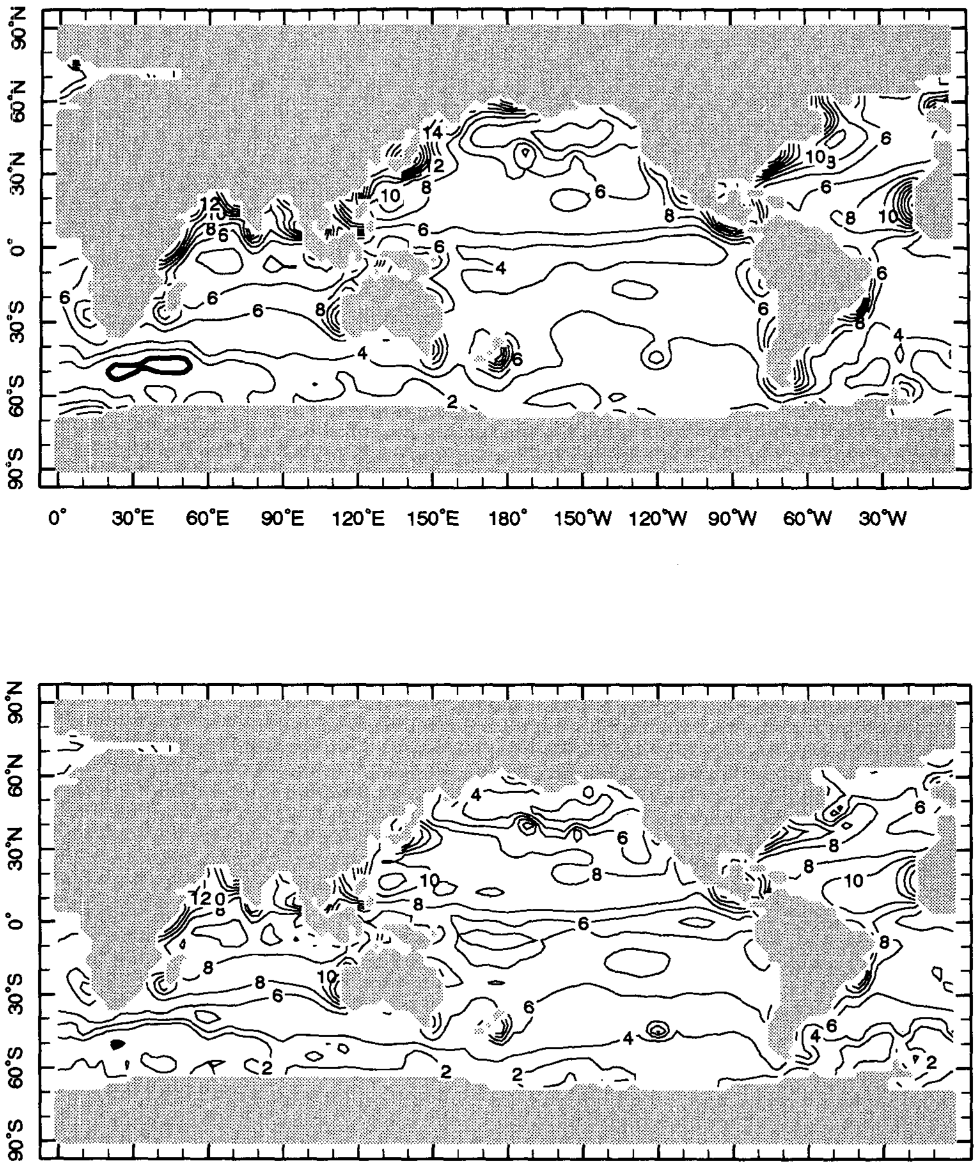


FIG. 1. Change of heat flux with SST (κ) for January (W m^{-2} K^{-1}). (a) Change in total heat flux, (b) contribution from change in latent heat flux, (c) contribution of sensible heat flux, and (d) contribution of net longwave radiative cooling.

by 2 K gives an estimate of the coupling coefficient κ. The atmospheric mixed layer model computes the changes in latent and sensible heat flux. However, the longwave radiative cooling also changes with SST and with the temperature and humidity of the mixed layer. To compute this we use the bulk formula for net longwave radiative cooling used by Esbensen and Kushnir (1981). The cloud cover needed in the formula is taken from Esbensen and Kushnir (1981), and the vapor pressure and temperature of the air are calculated by the atmospheric mixed layer model. As indicated above, we ignore any dynamical changes in the atmosphere that could possibly change the distribution of winds and surface solar radiation.

Since gradients of SST are unaltered, this case shows how the heat fluxes would change with SST in the absence of large changes in advection or diffusion. It should be compared to the estimates derived by Oberhuber (1988) who computed the sensitivity assuming the flux response was locally determined (note that he defined the heat flux as positive downward). It is also the sensitivity that is obtained for the largest possible

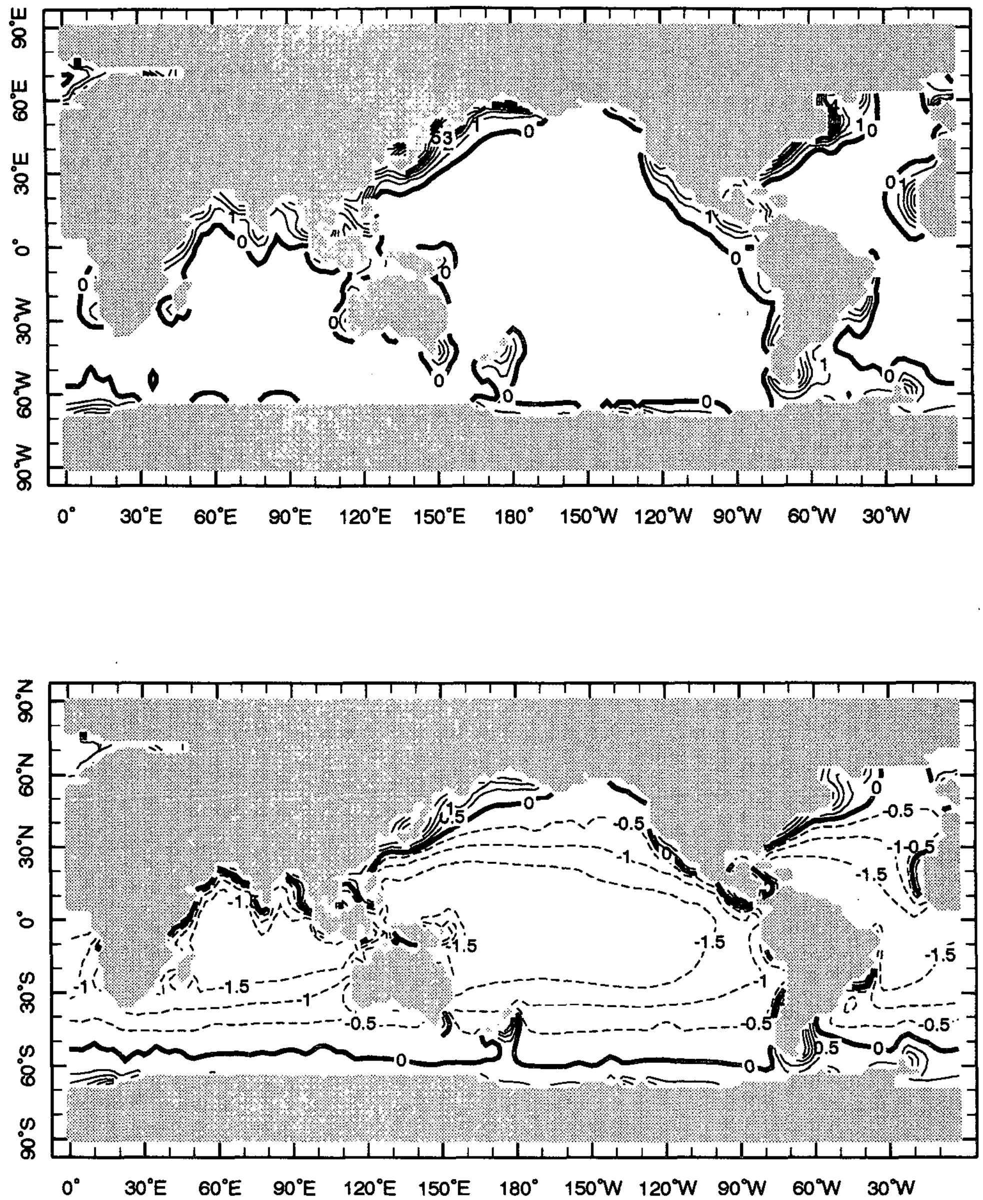
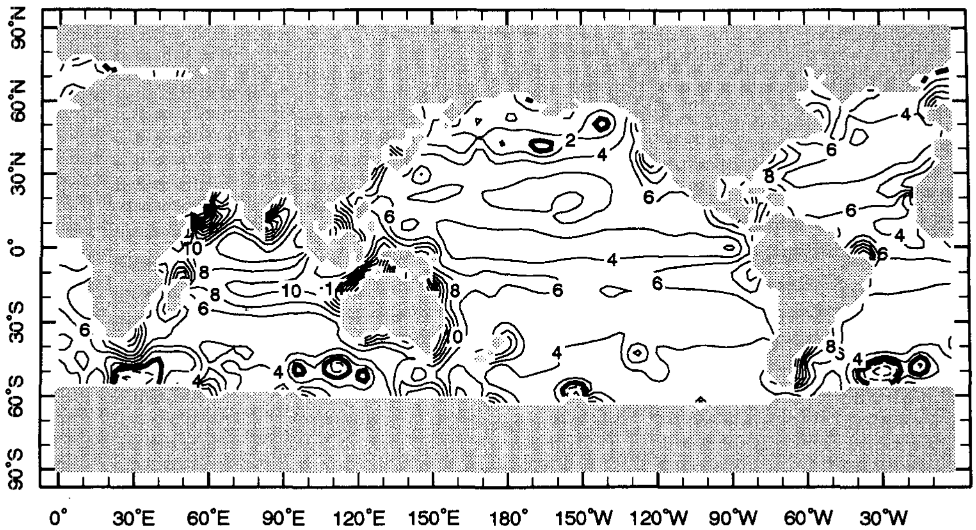


FIG. 1. (*Continued*)

spatial scale SST anomalies: a globally uniform warming or cooling of the planet.

Figures 1a and 2a show the model-derived estimates of κ for January and July. Typical open ocean values are in the range of 4 to 8 W m^{-2} K^{-1}, while values of up to 40 W m^{-2} K^{-1} can occur around the coasts. This spatial pattern is understandable in terms of the arguments presented by SBK. In open ocean regions the atmospheric mixed layer is in one-dimensional equilibrium. Here the thermodynamic properties of the atmospheric mixed layer easily adjust to the underlying SST so as to minimize the change in heat flux. There is a powerful negative feedback operating in which the lower levels of the atmosphere (which have a negligible heat and moisture content relative to the upper ocean) are forced to adjust to the underlying SSTs. Near the coasts there can be strong advection of air off the continents and the atmospheric mixed layer is not in equilibrium with the underlying ocean. This is especially so during winter. For example, in January there is a continual outflow of dry air from Asia and North America over warm waters offshore. The mixed layer is unable to come into equilibrium until it has advected some way offshore. Along its trajectory the fluxes are enhanced by positive SST perturbations giving large values of κ.



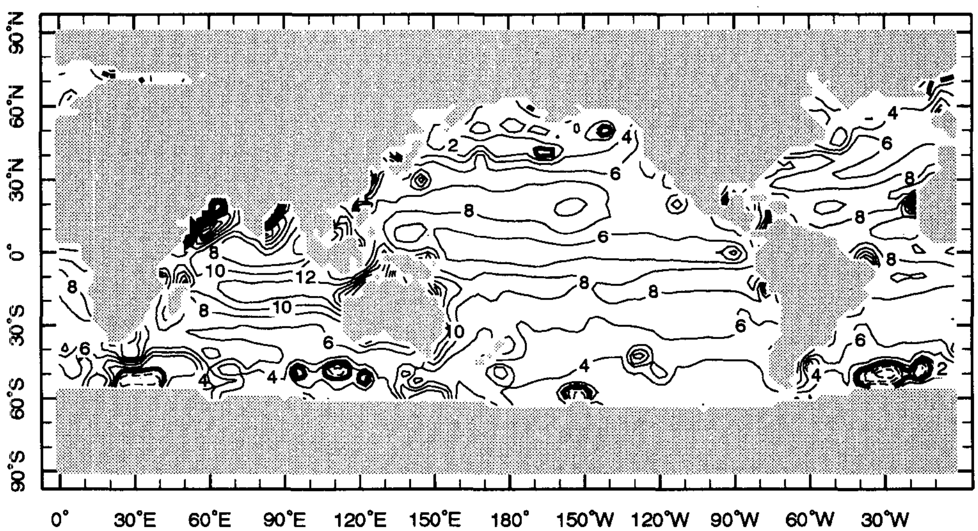


FIG. 2. As in Fig. 1 but for July.

Also in Figs. 1 and 2 we show the contributions to κ from the changes in latent heat, sensible heat, and net longwave cooling. The sensible heat is almost constant, indicating that the air temperature warms or cools by as much as the underlying SST. The latent heat flux increases by 6 or more W m^{-2} K^{-1} in open ocean regions with larger changes around the coast. The net longwave radiation loss decreases as the SST increases, except in regions of dry advection. This is because the mixed layer humidity increases, and the back radiation from the atmosphere to the surface increases, overwhelming the increase in upward flux at the surface. The longwave cooling is reduced by up to 2 W m^{-2} K^{-1}.

Clearly the sensitivity of the fluxes to changes in SST is much smaller than has been assumed to be the case in many ocean models and, instead, is of the magnitude suggested by RW. Betts and Ridgway (1989) present results for the changes in latent and sensible heat flux with SST in a tropical trade cumulus environment as calculated by a one-dimensional radiative–convective model. They find an increase of the latent heat flux by about 6 W m^{-2} K^{-1} and a near constant sensible heat flux. This is very similar to what we find.

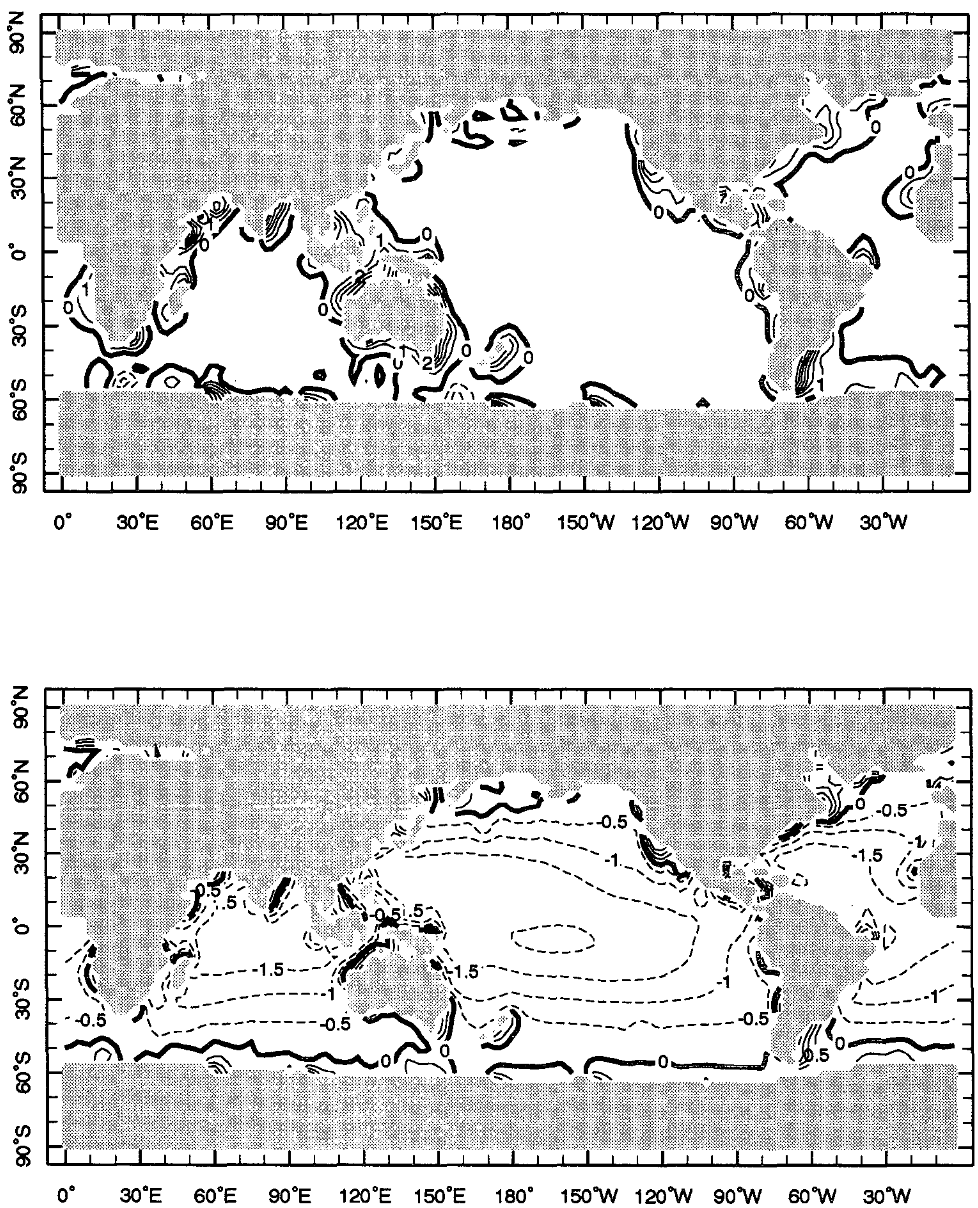


FIG. 2. (*Continued*)

Seager et al. (1988) subjected a tropical ocean model to a uniform change in surface heat flux and found the sensitivity to be a little larger than found here. This is largely attributable to the simpler heat flux formulation used in the earlier study. Also, the inclusion of ocean dynamics introduces additional spatial structure suggesting that in a fully coupled model, uniform changes in SST will not remain uniform as the coupled system evolves.

The reduced longwave cooling is familiar as an example of the "super greenhouse effect." For example, Inamdar and Ramanathan (1994) examined the radiative cooling of clear sky columns and found about a 5 W m^{-2} K^{-1} decrease in the net surface cooling with increasing SST. Our lower estimate may be the result of a reduction of this effect by the presence of clouds. That is, the net longwave radiation at the surface changes more with air humidity for clear sky conditions than for conditions where cloud absorption is dominant. The only regions where the net radiative cooling of the surface increases with SST are in regions of dry advection where the mixed layer moisture is not directly controlled by the SST. Longwave cooling of the surface is often subject to cursory treatment by ocean modelers (e.g., Seager et al. 1988; Philander and Siegel 1985). That it can reduce the flux sensitivity by one-

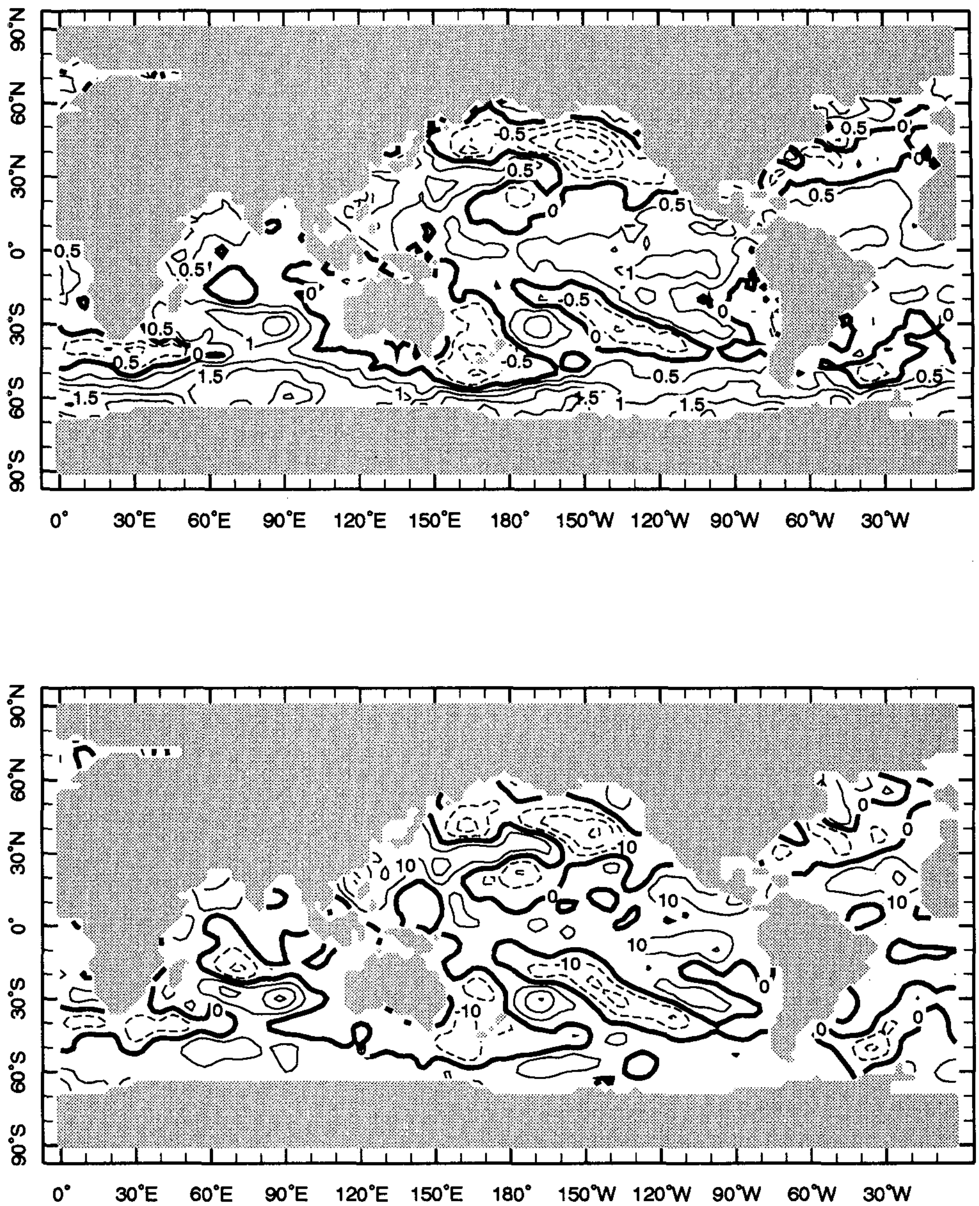


FIG. 3. (a) The observed SST anomaly (K) and (b) computed surface flux anomaly (W m^{-2} K^{-1}) for January 1988. The flux anomaly is the sum of the anomalies in latent, sensible, and longwave radiative fluxes for the case of no change in surface wind speed or cloudiness.

third strongly suggests that this term should in future receive more detailed treatment (e.g., Seager and Blumenthal 1994).

b. Sensitivity to observed SST changes

Here we compute the changes in model surface fluxes accompanying observed changes in SST. The results are presented as the change in total flux, rather than as a coupling coefficient, because regions of small or zero SST change prohibit calculation of an effective κ. We are not trying to address here the causes of the SST anomalies (they may even be forced by the atmosphere). Instead, we seek to quantify how, given a change in SST, the thermodynamics of the lower atmosphere respond in order to modify the surface heat flux. We are *only* concerned with the *thermodynamic* equilibrium response of the lower atmosphere and exclude consideration of heat flux anomalies that arise from changes in wind speed and solar radiation, both of which depend on changes in the circulation. Winds and cloudiness are kept fixed at their climatological values.

In Fig. 3a we show the global SST anomaly for January 1988, a time when there was a weak warm anomaly in the tropical Pacific and a cold anomaly in the North Pacific. Figure 3b shows the anomaly in total heat flux as calculated by the model. As expected, the heat flux increases over the warm tropical Pacific anomaly and decreases over the cold North Pacific anomaly. Other local anomalies in SST closely correlate with the heat flux anomalies. Comparing the size of the two fields, we see that the sensitivity is on the order of 10 W m^{-2} K^{-1} over the large tropical anomaly but can reach 20 W m^{-2} K^{-1} over the smaller-scale anomalies in the higher latitudes. In Figs. 4a and 4b we show maps of the SST and heat flux anomalies for January 1989, a time when there were cold SSTs in the tropical Pacific and warm SSTs in the North Pacific. The heat flux anomalies are reversed relative to January 1988, although the sensitivity is about the same.

The most striking difference between these results and the case of uniform changes in SST is the enhancement of the sensitivity. Clearly the sensitivity of fluxes to SST is very dependent on the scale of the SST anomaly. Imagine a small-scale positive SST anomaly. Increased evaporation will moisten the air above and reduce the flux. Yet, this will be counteracted by advection and diffusion of dryer air from surrounding regions with no SST anomaly. The larger the spatial scale of the SST anomaly, the less important are advection and diffusion, and the more locally determined is the sensitivity. It is in local equilibrium that the atmosphere adjusts to the maximum extent possible and minimizes the change in heat flux. Hence, the example in the previous section for a uniform change in SST gives the weakest sensitivity of fluxes to SST that is possible.

The scale dependence of the flux sensitivity has been previously emphasized by Bretherton (1982) and Kleeman and Power (1995), and our results agree with theirs. Even though the sensitivity is larger for regional SST anomalies, it is still well below that assumed in most ocean models. It is also slightly smaller than found by Kleeman and Power. The larger sensitivity in their case is explained by their assumption of a fixed relative humidity and specification of the temperature above the boundary layer at its climatological value, which allows less adjustment of near-surface temperature than in our more general case.

The implied timescale for damping of SST anomalies in the tropical Pacific is about 230 days. This is twice as long as assumed in the coupled model of the El Niño–Southern Oscillation introduced by Zebiak and Cane (1987). However, the 230-day value assumes a mixed layer 50 m deep. Although that is the depth of the mixed layer in the Zebiak and Cane model, the actual mixed layer depth in the tropical east Pacific is about half that. As the ocean mixed layer shoals, SST anomalies are damped faster by the SST–flux feedback. The shorter time assumed by Zebiak and Cane is designed to mimic this (S. Zebiak 1994, personal communication). Where the mixed layer is deeper, that is, over most of the tropical Pacific, the damping time assumed by Zebiak and Cane (1987) is probably a factor of 2 too small.

Our estimates agree quite well with the tropical Pacific values of Barnett et al. (1991) derived from an atmospheric general circulation model. Their estimate is, however, for the change in total flux, including the effects of variations in winds and cloud cover. The contribution of the latter two effects, which involve dynamical responses to SST anomalies, deserves to be assessed in the future.

5. Discussion of implications

We have presented evidence, based on modeling of the atmospheric mixed layer, that the sensitivity of the surface heat fluxes to SST is much smaller than typically assumed. The mixed layer thermodynamics are inevitably forced to adjust to the changed SST because the atmosphere can store only small amounts of heat and moisture relative to the ocean. The most commonly used thermal boundary conditions instead implicitly assume the atmosphere has an infinite heat capacity and that its moisture content is independent of the underlying SST.

Though the lower values of the coupling coefficient are close to those of RW, the flux formula implied here still differs from theirs. We can cast our surface heat flux in their form (minus the diffusion term) by linearizing the flux around the modeled mean flux:

$$Q = Q_0 + \kappa(T - T_0), \tag{5}$$

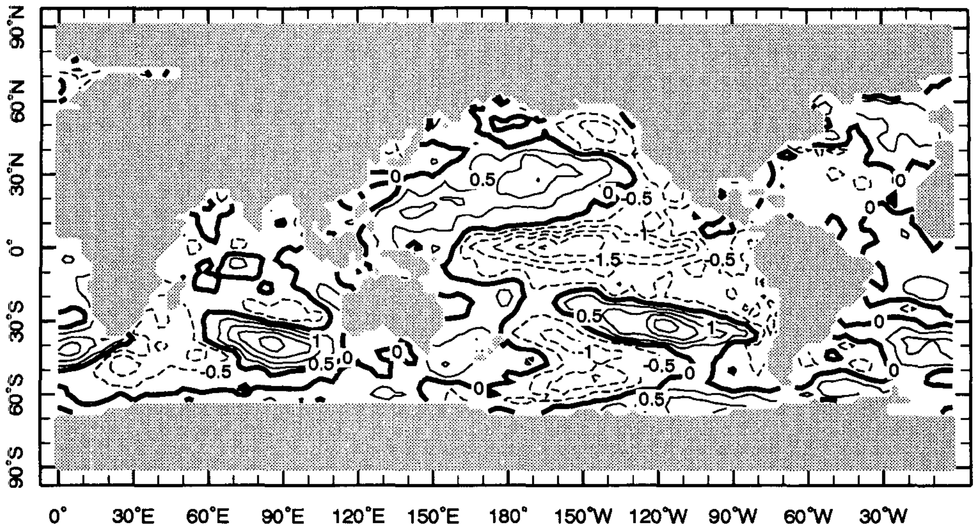
where Q_0 is the net surface heat flux for surface temperature T_0, κ is the coupling coefficient derived numerically by the experiments with a uniform change in SST, and T is the perturbed ocean temperature. Defining a reference temperature T_r, we can rewrite this as

$$Q = \kappa(T - T_r), \tag{6}$$

with

$$T_r = T_0 - Q_0/\kappa. \tag{7}$$

Rahmstorff and Willebrand use this form with a small κ but take T_r to be within a few kelvin of the observed surface temperature. We calculated T_r using the mean climatological fluxes of latent, sensible, and longwave fluxes derived from the model, and the satellite-observed climatological net solar radiation described by Li and Leighton (1993). It is shown for January and July in Fig. 5. What is most noticeable is that in regions of strong downward heat fluxes (Q_0 negative) T_r is much greater than T_0 and reaches a maximum of more than 330 K in the equatorial Pacific cold tongue. At high latitudes, where the net heat flux is upward and large, T_r is much less than T_0. The meridional gradient of T_r is considerably greater than that of the observed SST.



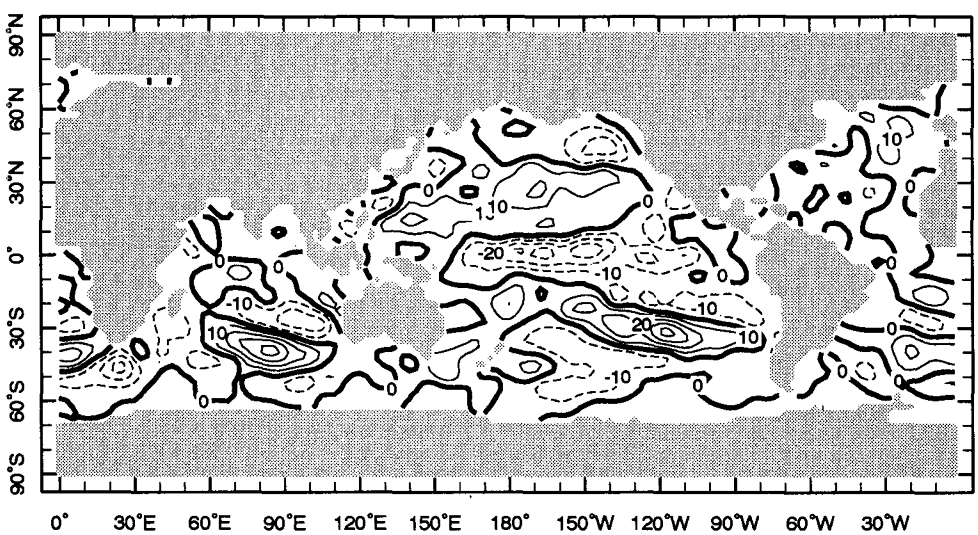


FIG. 4. As for Fig. 3 but for January 1989.

What does T_r represent? It is often thought to be the temperature the ocean needs to adopt for the net surface heat flux and ocean heat transport to be zero (e.g., RW). But this is a quite misleading interpretation because, as calculated here, it incorporates information from the current state of affairs in which the ocean *does* transport heat. Consider the tropical east Pacific. The downward heat flux here is large because ocean dynamics cool this region so much. Hence, from (7), the restoring temperature must be large. This T_r is then the temperature the ocean would obtain if the ocean dynamics were shut off, *but the atmosphere continued to heat this region as if the cooling were still there.* Essentially the linearization breaks down. In previous work we have shut off the dynamics and calculated the SST that results and, as expected, the east Pacific attains an SST similar to the west Pacific (Seager et al. 1988).

The restoring temperature as defined by (7) has little physical relevance. However, that does not mean it can be replaced with a value that is close to the observed SST since it is still the T_r that needs to be used in combination with κ if the correct flux is to be represented in the Newtonian cooling form of (6). We are not suggesting that anyone use the T_r derived here since its use with an erroneous ocean model would likely lead to

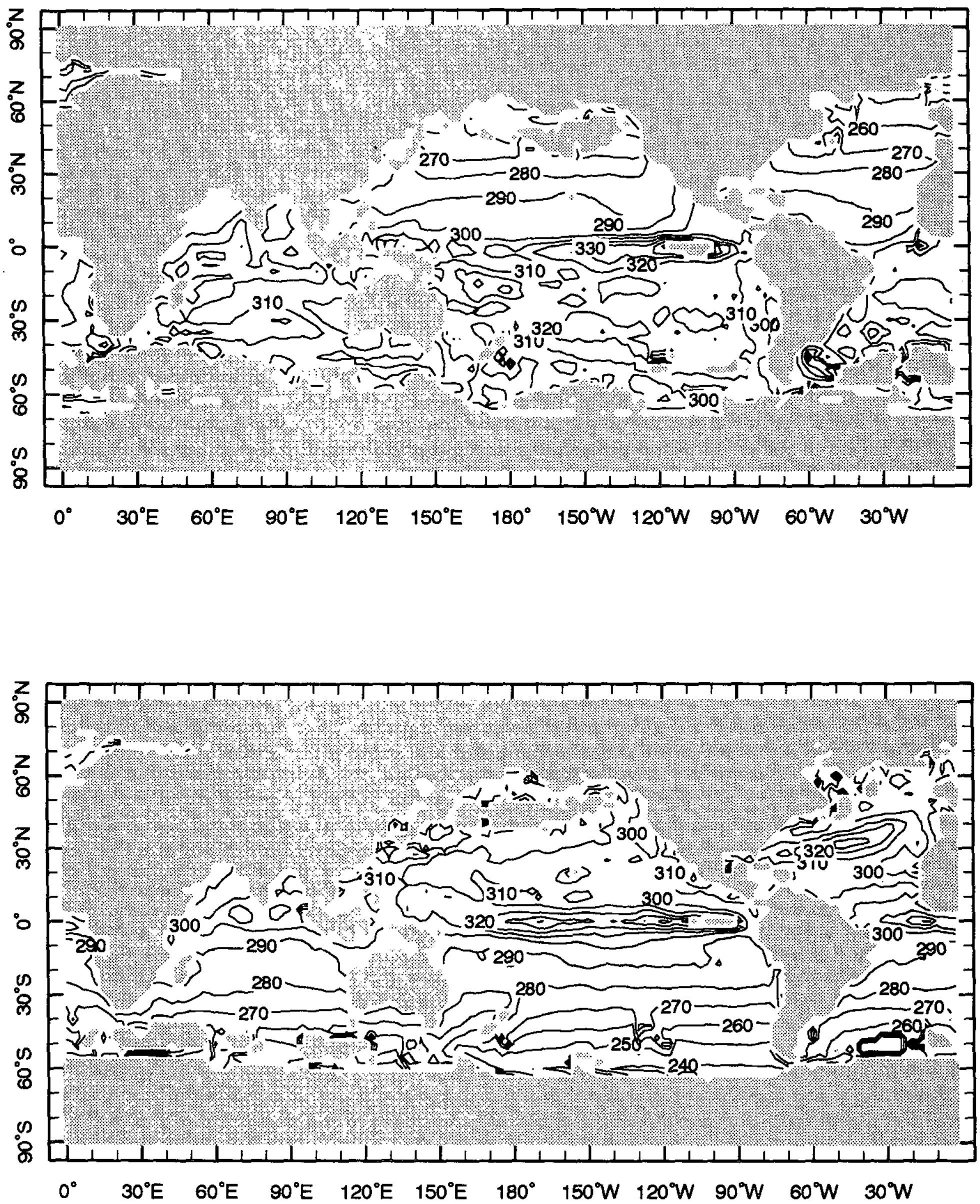


FIG. 5. The implied restoring temperature (K) calculated by linearization of the surface heat flux around its mean value for (a) January and (b) July. Some very large values have been blacked out for ease of presentation.

large SST errors. The only proper procedure is to explicitly include the feedbacks between SST and fluxes.

It has been shown, in agreement with Kleeman and Power (1995), that the flux sensitivity to SST changes is very dependent on the scale of the SST anomaly. The larger the scale of the anomaly, the more locally determined is the response, and the smaller the sensitivity. It is also possible for an SST perturbation of one sign to induce flux perturbations of the opposite sign around its flanks due to the effects of advection and diffusion. Clearly, the flux response is inherently nonlocal and scale dependent.

6. Conclusions

Recent modeling studies of the thermohaline circulation have used a damping on SST anomalies that far exceeds what is realistic (e.g., Marotzke and Willebrand 1991; Weaver and Sarachik 1991a; Weaver et al. 1993). The large values were based on the assumption that the atmosphere did not adjust to changes in SST that, given the small heat and moisture storage of the atmosphere relative to the ocean, is a particularly poor assumption. We have used a model of the lower atmosphere to assess the implied sensitivity. For a uniform change in SST we derive small values of the order of 4 W m^{-2} K^{-1} in open ocean regions. The latent heat flux increases with SST and is offset, to some extent, by a reduction in longwave radiative cooling of the surface.

The sensitivity of the flux to changes in SST increases as the scale of the SST anomaly decreases. The sensitivities derived for a uniform change represent the minimum possible. As the scale of the SST anomaly is reduced, the response becomes more nonlocal, and the atmospheric mixed layer is unable to come into local equilibrium with the underlying SST, enhancing the sensitivity. For observed SST anomalies we found the sensitivity could increase to around 15 W m^{-2} K^{-1}. This is still well under half the values typically assumed. What is more, the scale dependence and nonlocality of the response cannot be captured by a simple restoration of SST to a prescribed value even if the coupling coefficient is spatially variable.

Rahmstorf and Willebrand (1995) found that reduced damping of SSTs will affect the variability of modeled thermohaline circulations. Weaver et al. (1993) have also speculated on the sensitivity of modeled circulations to assumptions regarding the surface heat fluxes. In present models, salinity forcing can result in changes in column stability that induce changes in the circulation. The altered circulation will bring more or less water poleward, but this will not effect column stability because the SST is rapidly adjusted back to a fixed value by the strong restoring conditions. The ocean's role in transporting heat is effectively disabled. In contrast, if the SST could evolve, then increased overturning would advect in warmer water from the subtropics, which would increase the buoyancy of high-latitude water and reduce the overturning. Allowing the SST to evolve in a reasonable way introduces a negative feedback between the circulation and the surface fluxes. The implications of this need to be explored with models that contain more realistic surface boundary conditions.

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